Doctoral Thesis

Katabatic wind over Greenland
Comparison between model results and observations

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Publication Date: 1999

Permanent Link: https://doi.org/10.3929/ethz-a-002094852

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KATABATIC WIND OVER GREENLAND:
Comparison between model results and observations

A dissertation submitted to the
Swiss Federal Institute of Technology, Zürich

for the degree of Doctor of Natural Sciences

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Zürich, 1999
Acknowledgments

I am indebted to many people who supported me in the course of this study and during the field work on the Greenland ice sheet.

First of all, I would like to thank my advisor Prof. Atsumu Ohmura. He gave me the opportunity to join the Greenland project, which was for me a great personal and scientific experience. I will always keep good memories of the time I spent in Greenland.

Special thanks go to Heinz Blatter, Jann Forrer, Pierluigi Calanca, Karl Schroff, Marcel Haefliger from the ETH, and to Konrad Steffen, Julienne Stroeve and Waleed Abdalati from the University of Colorado, Boulder. They have all been my fellow researchers on the ice sheet.

I am extremely grateful to Prof. Thomas Parish who gave me the opportunity to stay one year at the University of Wyoming. During this time I learned a lot not only about the model but also about the way of life in the Wild West.

At the Institute of Geography, a lot of people, too numerous to mention all of them individually, helped me in one way or another and contributed to a pleasant and stimulating environment.

I would also like to thank my wife Angela who always gave me moral support in the course of this work.

The European Center for Medium-Range Weather Forecasts is acknowledged for providing the analyses. The ETH Greenland expedition was financed by the Swiss Federal Institute of Technology (ETH), Zürich (Grant No. 0-20-013-90 and 0-40-040-90) and the Swiss National Science Foundation (Grant No. 21-27449-89, 20-32649-91, 20-36396-91 and 20-39129-93).
Abstract

The aim of this study is to develop a mesoscale model which is able to simulate the katabatic wind for the entire Greenland ice sheet for any season of the year. The performance of the model will be validated with measured data.

The Institute of Geography, ETH Zurich carried out a scientific expedition during the summer seasons 1990 and 1991 on the west slope of Greenland. In this area of Greenland the katabatic wind is usually present. Part of this project was to investigate the mechanism of the katabatic wind with continuous measurement of wind speed and wind direction within the surface layer. Twice a day weather balloons measured the wind profile of the entire troposphere. After the two summer seasons an automatic weather station was installed. This station has been continuously measuring the temperature, wind speed, and wind direction from September 1991 until today. In the meantime more automatic weather stations have been installed in different places on the Greenland ice sheet. For the present work measurements from four different stations have been used, with the ETH/CU-Camp on the west slope of the ice sheet, Humboldt on the north-west slope, Tunu on the north-east slope of Greenland, and GISP2 near the summit. The first three stations are situated within the katabatic wind regime while the last station is not influenced by the katabatic wind.

First, a model developed at the University of Wyoming, which had been tested for the katabatic wind in the Antarctic, mainly for winter situation, has been transformed to the situation in Greenland. It was mainly the topography and the Coriolis term which had to be adapted. Despite the fact that Greenland has a different position in respect to the pole than Antarctica, the transformation of the model was successful. A comparison between measurements and model results for all three stations on the slope correspond well in respect to wind speed, wind direction, and temperature. Near the summit the difference between measurements and model results is larger, especially for the wind speed. This, however, is not surprising since in this area the influence of synoptic phenomena becomes important. These phenomena have not been treated in this model.

In a second step, the model was extended to simulate the katabatic wind for summer conditions. Due to the large influence of the solar radiation in summer, all energy terms had to be modified. The parameterization of the energy terms was determined by measurements from the ETH/CU-Camp. The albedo turned out to be the critical parameter for the short wave radiation. Accurate values for the albedo made it possible to have better simulations of the short wave radiation. For the long wave radiation melting or freezing processes on the snow surface are important. Another significant factor is the humidity of the air and thus the clouds. All these factors have been treated in a very simplified manner in the present model. This leads to differences between the measured and modelled results for long wave radiation and for temperature. For the turbulent fluxes the difference between model and measurements comes mainly from the limited information on the surface roughness length.

Despite all the simplifications the model is able to perform a satisfactory simulation of the katabatic wind for winter and for summer conditions. Especially the model wind direction
corresponds very well with observed values, while the wind speed is less well simulated. The energy terms are well enough parameterized to enable the simulation of the katabatic wind for Greenland in summer and winter and on any place on the ice sheet.
Zusammenfassung


In einem ersten Schritt wurde ein bestehendes Modell von der University of Wyoming, welches für Wintersimulation des katabatischen Windes in der Antarktis entwickelt wurde, so geändert, dass es für die Wintersimulation in Grönland funktioniert. Es waren vor allem die Topographie und die Corioliskraft welche angepasst werden mussten. Trotz der unterschiedlichen Lage der beiden Eisschilder in Bezug zum Pol war die Transformation erfolgreich. Der Vergleich mit den Messwerten an allen drei Stationen, welche im katabatischen Windregime liegen, ergab eine gute Übereinstimmung bezüglich Windgeschwindigkeit, Windrichtung und Temperatur. Im Bereich von Zentralgrönland ist vor allem die Übereinstimmung der Windgeschwindigkeit weniger gut, was aber weiter nicht erstaunlich ist, da in diesem Gebiet synoptische Einflüsse den Wind bestimmen. Diese wurden im Modell nicht mit berücksichtigt.

zurückzuführen. Trotz gewisser Vereinfachungen im Modell führt die Simulation für eine Situation, welche den Bedingungen im Juli entspricht, zu recht guten Resultaten, vor allem für die Windrichtung und etwas weniger gut für die Windgeschwindigkeit. Das Muster des Windfeldes entspricht auf jeden Fall den Beobachtungen.
Contents

Acknowledgments .......................................................... i
Abstract ........................................................................... ii
Zusammenfassung ............................................................ iv
Contents .............................................................................. vii
List of Figures ....................................................................... ix
List of Tables ......................................................................... x

1 Introduction ........................................................................ 1

2 The Model ............................................................................ 3

  2.1 Governing equations ...................................................... 3
      2.1.1 Coordinate transformation ........................................ 3
      2.1.2 Grid structure ........................................................ 5
      2.1.3 Time integration ..................................................... 5
  2.2 Parameterization of physical processes ............................. 6
      2.2.1 Longwave radiation ............................................... 6
      2.2.2 Shortwave radiation ............................................... 8
      2.2.3 Surface energy budget and surface temperature .......... 10
      2.2.4 Planetary boundary layer formulation ....................... 11
      2.2.5 Horizontal diffusion .............................................. 13
  2.3 Boundary conditions .................................................... 14
      2.3.1 Lateral boundary conditions .................................... 14
      2.3.2 Top of the model .................................................. 14

3 Field observations ............................................................ 16

  3.1 ETH/CU Camp ............................................................. 17
      3.1.1 Wind measurements .............................................. 17
      3.1.2 Temperature measurements .................................... 22
      3.1.3 Energy flux ......................................................... 25
  3.2 Humboldt and Tunu ........................................................ 26
  3.3 GISP2 ............................................................................. 31
CONTENTS

4 Winter Simulation 34
  4.1 Introduction ............................................. 34
  4.2 Initial conditions ........................................ 34
  4.3 Wind ....................................................... 36
  4.4 Temperature ............................................... 44
  4.5 Energy terms .............................................. 50

5 Summer simulation 52
  5.1 Introduction ............................................... 52
  5.2 Initial conditions ........................................ 52
  5.3 Wind ....................................................... 55
  5.4 Temperature ............................................... 61
  5.5 Energy terms at the surface at the ETH/CU Camp .......... 65
    5.5.1 Shortwave radiation .................................. 65
    5.5.2 Longwave radiation ................................... 67
    5.5.3 Sensible heat flux .................................... 70

6 Discussion and Conclusion 74

A List of Symbols 76

B Coordinate transformation 79
  B.1 Basic atmospheric equations ................................ 79
  B.2 Generalized vertical coordinate system ..................... 79

References 84
### List of Figures

2.1 Some parts of the staggered grid used in the model .................................................. 5
3.1 A map of Greenland showing the position of the automatic weather stations .......................... 16
3.2 Mean daily wind speed cycle from a ten-meter level at the ETH/CU Camp .......................... 20
3.3 Wind-roses for 1995 at the ETH/CU Camp ................................................................. 21
3.4 Mean monthly ice temperature profile at the ETH/CU Camp ........................................... 23
3.5 Mean daily temperature cycle from a ten-meters level at the ETH/CU Camp ....................... 24
3.6 Wind-roses for 1996 at Humboldt .................................................................................. 28
3.7 Wind-roses for 1996 and 1997 at Tunu ............................................................................ 30
3.8 Wind-roses for 1994 at GISP2 ....................................................................................... 33
4.1 The original terrain computed with data from the Alfred Wegener Institute ......................... 35
4.2 The smoothed terrain used in the 3-d model ....................................................................... 35
4.3 Wind vectors for the first sigma level for winter simulation ............................................... 37
4.4 Comparison of the mean daily wind speed at four different locations for the winter simulation .......................................................... 38
4.5 Comparison of the mean daily wind direction at four different locations for the winter simulation .................................................................................................................. 39
4.6 Wind vector field representing the January model run for the north-western slope of Greenland .......................................................................................................................... 40
4.7 Cross section of the wind at north west Greenland ............................................................ 41
4.8 Cross section of down slope wind speed component at the latitude of the summit ............... 42
4.9 Cross section of cross slope wind speed component at the latitude of the summit ............... 43
4.10 Wind profiles at north west and central Greenland ........................................................... 43
4.11 Mean daily surface temperature for January ...................................................................... 44
4.12 Comparison of the mean daily temperature at four different locations for the winter simulation .................................................................................................................. 45
4.13 Cross section for the potential temperature at 72 degrees north for January ....................... 46
4.14 Profiles of potential temperature for the west slope and for north-west Greenland ............. 47
4.15 Potential temperature field at the lowest $\sigma$-level resulting from the January model run for the north-western slope of Greenland ........................................................................... 48
4.16 Cross section for the potential temperature in north-western Greenland .......................... 49
4.17 Energy terms from the 72 h simulation for January .......................................................... 51
LIST OF FIGURES

5.1 Surface albedo for the July simulation ........................................ 54
5.2 Sea surface temperature for the July simulation .............................. 54
5.3 Wind field at night for July simulation ........................................ 55
5.4 Wind field in the afternoon for July simulation ............................... 56
5.5 Wind speed and wind direction for the month of July at the ETH/CU Camp 57
5.6 Wind speed and wind direction for July at Humboldt .......................... 58
5.7 Wind speed and wind direction for the month of July at Tuni ............... 58
5.8 Wind profiles at ETH/CU Camp .................................................. 59
5.9 Cross section of the x-component of the wind at the latitude of the ETH/CU Camp .................................................. 60
5.10 Cross section of the y-component of the wind speed at the latitude of the ETH/CU Camp .................................................. 60
5.11 Mean daily surface temperature for the Greenland ice sheet ............... 61
5.12 Mean daily temperature cycle at ETH/CU Camp and Humboldt for July . 62
5.13 Mean daily temperature cycle at Tuni and GISP2 for the month of July 63
5.14 Mean monthly temperature profile at the ETH/CU Camp for the month of July .................................................. 63
5.15 Cross section of temperature at the latitude 72 degrees North for the month of July .................................................. 64
5.16 Shortwave radiation at ETH/CU Camp for the month of July .............. 65
5.17 Shortwave net radiation for the Greenland ice sheet .......................... 66
5.18 Comparison of longwave radiation at ETH/CU Camp for July .............. 67
5.19 Longwave net radiation for the Greenland ice sheet .......................... 68
5.20 Sensible heat flux at ETH/CU Camp ........................................... 70
5.21 Sensible heat flux for the Greenland ice sheet ................................ 71
5.22 Latent heat flux at ETH/CU Camp ............................................. 72
5.23 Latent heat flux for the Greenland ice sheet .................................. 73
List of Tables

3.1 Monthly mean wind speed on the ETH/CU Camp .......................... 18
3.2 Monthly directional constancy of the wind at the ETH/CU Camp ........ 19
3.3 Monthly mean air temperature at the ETH/CU Camp ....................... 23
3.4 Monthly mean values of the radiation components at the ETH/CU Camp . 25
3.5 Monthly mean air temperature, wind speed, directional constancy, incoming shortwave radiation, reflected shortwave radiation, net radiation, and Albedo at Humboldt ........................................... 27
3.6 Monthly mean air temperature, wind speed, directional constancy, incoming shortwave radiation, reflected shortwave radiation, net radiation, Albedo at Humboldt ........................................... 29
3.7 Monthly mean air temperature at GISP2 .................................... 31
3.8 Monthly mean wind speed and directional constancy at GISP2 .......... 32
5.1 Albedo from stations in Greenland ............................................. 53
1 Introduction

The possibility of a sea level rise as a consequence of the anticipated anthropogenetic greenhouse warming (Warrick and Oerlemans 1990) has led to extensive scientific activities on the Greenland ice sheet since 1990 (Abdalati, 1996; Stearns et al., 1994; Oerlemans and Vugts, 1993; Ohmura et al., 1992; Ohmura et al., 1991). These activities have been motivated by the general recognition of the importance of the Greenland ice sheet in the case of a global warming. The climatology section of the Department of Geography, ETH Zurich conducted a scientific expedition during the summer seasons 1990 and 1991. The main topic was the energy and mass balance of the Greenland ice sheet. The present work has been done within the scope of this project. It treats the large scale wind field above the surface.

The Greenland ice sheet is one of the largest in the world, apart from the Antarctic ice sheet, and the only one in the northern hemisphere. It has a typical shield shape climbing at first steeply from sea level to 1000 m and then gently to the highest elevation of 3230 m. Greenland stretches from 60°N to 83°N and forms together with Asia and North America an alternation of land and ocean. Next to many similarities between the Greenland and Antarctic ice sheets there is a main difference which is important for the present work. The tropospheric circulation at the latitude of Greenland is dominated by the circumpolar vortex. With its traveling disturbances the vortex influences the katabatic wind in a much stronger form than in Antarctica.

One of the main features of the lower atmosphere over an ice sheet is the persistent down-slope or katabatic wind, which influences the energy balance of the ice sheet considerably. The katabatic wind is a gravity-driven atmospheric current that is forced by the cooling of air adjacent to a sloped surface. It is generally classified as a mesoscale phenomenon. The exception to this rule arises when the length scale of the slope becomes as large as the scale of a continent and when the surface cooling conditions required for the production of a strong sloped inversion are fulfilled. This is the case at least for Greenland and Antarctica, especially during the long polar night.

Currently, the knowledge about the katabatic wind was mostly acquired from researches done for Antarctica. In the past, only few investigations were made for the Greenland ice sheet. The scientific investigations of the katabatic wind started after the Second World War. Fleagle (1950) explained the downward flow with a two layer model. Ball (1956) described the phenomenon observed along the coast of Antarctica of a sudden decay of katabatic wind before it passed the coastline. According to Ball, an hydraulic jump of the cold surface layer is the cause for this decay. Manins and Sawford (1979) have shown that the mixing of a cooled air layer with the surrounding air is an important retarding stress of the katabatic flow. The cause for the strong katabatic wind at Cape Denison and Port Martin was described by Parish (1981, 1982, 1984). He demonstrated that large scale wave-like features in the interior of the ice sheet cause a marked convergence of cold air upstream off the coast. Several studies with model simulations were made to understand the phenomenon of the katabatic wind (Parish, 1983; Waight, 1987; Parish and Bromwich, 1991; Gallée and Schayes, 1992). All of them treat the conditions in Antarctica. Recently, first simulations with two-dimensional models were made for Greenland (Gallée et al.,
1 INTRODUCTION

The purpose of this work is

- to introduce a three dimensional model simulating the katabatic wind for the entire Greenland ice sheet
- to make the model applicable for any time of the year
- to discuss a winter simulation by comparing model temperature, wind speed, and wind direction with measurements from automatic weather stations
- to discuss a summer simulation by comparing temperature, wind speed, wind direction, and energy terms with measurements from automatic weather stations and measurements carried out during two field campaigns.

This work relies on a model developed at the University of Wyoming (Parish, 1984; Cerni and Parish, 1984; Waight, 1987; Parish and Waight, 1987). This model is based on a parameterization of the dynamic processes made by Anthes and Warner (1978). Since the model has been used only for winter conditions in Antarctica, part of it had to be modified. The topography of Greenland had to be introduced and transferred from the southern hemisphere to the northern hemisphere. Some of the energy terms have been changed. First, the short wave radiation and turbulent flux parameterization were replaced. Then, snow and surface parameters were modified according to the knowledge gained during the ETH Greenland Expedition (Ohmura et al., 1991; Ohmura et al., 1992).
2 The Model

2.1 Governing equations

The motions of the atmosphere are governed by fundamental laws of physics. Consequently, the model consists of prognostic equations for momentum, temperature, and mass (continuity) and of two diagnostic equations (hydrostatic and vertical velocity) (Holton 1979).

In this version of the model, humidity is treated pseudodiagnostically. The relative humidity is taken from a given three-dimensional distribution (constant in time) from the European Center of Medium-Range Weather Forecasts (ECMWF). The specific humidity is computed through the saturation vapor pressure from the temperature and pressure, which are prognostic variables. The model also includes a prognostic equation for the surface temperature.

2.1.1 Coordinate transformation

As typical for mesoscale models, the equations are written in terrain following $\sigma$-coordinates. In this specific case, the system proposed by Anthes (1972) is used:

$$ p = \frac{p_s - p_t}{p_s - p_i} = \frac{p - p_i}{p_s} \text{ and } p_* = p_s - p_t. \quad (2.1) $$

The coordinate transformation from the Cartesian coordinates to the $\sigma$-system of each governing equation is shown in detail in appendix B. In this system $p_t$ is the pressure at the top of the model, $p_s$ the pressure at the surface, and $p$ the local pressure, and the prognostic and diagnostic equations have the following form:

**Continuity equation**

The continuity equation is

$$ \frac{\partial p_*}{\partial t} = -\nabla_\sigma (p_* \vec{v}) - \frac{\partial p_* \sigma}{\partial \sigma}, \quad (2.2) $$

where $t$ is the time, $\sigma$ the vertical component of the velocity, $\vec{v}$ the horizontal wind speed, and

$$ \nabla_\sigma = \frac{\partial}{\partial x} + \frac{\partial}{\partial y} \quad (2.3) $$

in the $\sigma$-coordinate system. Integration from the surface to the top of the model leads to

$$ \frac{\partial p_*}{\partial t} = -\int_0^1 \left( \frac{\partial p_* u}{\partial x} + \frac{\partial p_* v}{\partial y} \right) d\sigma. \quad (2.4) $$
Equations of motion

The horizontal equation of motion is

\[
\frac{di}{dt} + f(k \times \vec{v}) = -\frac{RT}{p_* + \frac{\mu}{\sigma}} \nabla \Phi - \nabla \Phi + F, \tag{2.5}
\]

where \( T \) is the temperature, \( f \) is the Coriolis parameter, \( R \) is the gas constant for dry air, and \( \Phi \) is the geopotential. The addition of Eq.2.2 multiplied with \( \tilde{v} \), a diffusion term, and multiplying with \( p_* \) leads to

\[
\frac{dp_*}{dt} + \frac{dp^*}{dt} - \frac{dp^*}{dt} + f_p^* v - p_* \left( \frac{RT}{p_* + \frac{\mu}{\sigma}} \frac{dp_*}{dx} + \frac{\partial \Phi}{\partial x} \right) + p_* F_u + F_{Uh}, \tag{2.6}
\]

\[
\frac{dp^*}{dt} + \frac{dp^*}{dt} - \frac{dp^*}{dt} - f_p^* v - p_* \left( \frac{RT}{p_* + \frac{\mu}{\sigma}} \frac{dp_*}{dy} + \frac{\partial \Phi}{\partial y} \right) + p_* F_v + F_{Vh}, \tag{2.7}
\]

where \( u \) and \( v \) are the horizontal components of the velocity. The first three terms on the right hand side of Eqs. 2.6 and 2.7 describe the advection written in flux form. The fourth term describes the influence of the earth’s rotation (Coriolis effect). The fifth term in both equations is an expression of the pressure gradient force. \( F_u \) and \( F_v \) represent the friction, whereas \( F_{Uh} \) and \( F_{Vh} \) account for the horizontal diffusion necessary for numerical stability (Waight 1987).

Thermodynamic energy equation

The thermodynamic equation is

\[
\frac{dT}{dt} = -u \frac{dT}{dx} - v \frac{dT}{dy} - \sigma \frac{dT}{\sigma} + \frac{RT\omega}{c_p} \frac{\partial Q}{\partial \Phi} + F_{Tv} + F_{Th}, \tag{2.8}
\]

where \( c_p \) is the heat capacity of dry air at constant pressure, \( Q \) the radiative flux, \( F_{Tv} \) the effect due to the sensible heat flux, \( F_{Th} \) the effect due to the horizontal sub-grid scale diffusion processes necessary for numerical stability, and \( \omega \) the change of pressure with time of a parcel given by

\[
\omega = p_* \sigma + \sigma \left( \frac{\partial p_*}{\partial t} + u \frac{\partial p_*}{\partial x} + v \frac{\partial p_*}{\partial y} \right). \tag{2.9}
\]

Hydrostatic equation

Mahrt (1982) demonstrates that, the vertical component of the equation of motion reduces to the hydrostatic equation, if the slope is sufficiently small,

\[
\frac{\partial \Phi}{\partial \ln \left( \sigma + \frac{\mu}{p_*} \right)} = -RT. \tag{2.10}
\]
Equation for the vertical velocity

The equation for the vertical velocity results from integration of the mass continuity equation:

\[
\hat{\sigma} = \frac{1}{p^*} \int_{1}^{\sigma} \left( \frac{\partial p^* u}{\partial x} - \frac{\partial p^* v}{\partial y} - \frac{\partial p}{\partial t} \right) d\sigma.
\]

(2.11)

2.1.2 Grid structure

Anthes (1972) showed that a staggered Arakawa grid gives the best results. In the horizontal plane such a grid consists of two sets of prediction points, one for the horizontal velocity and one for all the other variables. In the vertical planes all variables are calculated on the \(\sigma\)-levels except for the vertical velocity, which is calculated for a level between two \(\sigma\)-level. In the horizontal planes the horizontal velocities are calculated between the grid points, where also all the other components are calculated (see Fig. 2.1).

![Diagram of grid structure](image)

Figure 2.1: Some parts of the staggered grid used in the model. The left figure shows a horizontal plane and the right figure a vertical plane. The temperature, geopotential, and pressure terms are defined on the grid, the velocity terms on the levels in-between.

2.1.3 Time integration

The time integration scheme used in the model is the one developed by Shuman and Hovermale (1968) and generalized by Brown and Campana (1978). This particular scheme allows larger time steps than the conventional leapfrog scheme, however, it produces almost identical results. To achieve stability of the scheme one has to compute the values of \(p^*\) and \(\Phi\) before computing the momentum values.

In addition, a frequency filter is applied which damps high frequency oscillations. The strength of smoothing can be controlled by setting a parameter \(\beta\), where a value of \(\beta = 1\)
causes strong damping and a value of $\beta = 0$ deactivates the filter (Waight 1987). For the simulation presented in this work the value was set $\beta = 0.1$.

2.2 Parameterization of physical processes

2.2.1 Longwave radiation

Radiative cooling of the lowest layer is an important process for the development of the katabatic wind. In the model an approach described in Liou (1980) and in Cerni and Parish (1984) is used. The total upward and downward infrared flux density can be expressed by

$$ F^\uparrow(\Upsilon) = \int_0^\infty \pi B_\nu (T_S) T_\nu(\Upsilon) d\nu + \int_0^\infty \int_0^\Upsilon \pi B_\nu [T(\Upsilon')] dT_\nu(\Upsilon - \Upsilon') d\nu $$

and

$$ F^\downarrow(\Upsilon) = \int_0^\infty \int_0^\Upsilon \pi B_\nu [T(\Upsilon')] dT_\nu(\Upsilon' - \Upsilon') d\nu, $$

where $T_\nu$ denotes the monochromatic slab transmission function at frequency $\nu$. $F^\uparrow$ represents the upward and $F^\downarrow$ the downward longwave radiation flux density; the unit is Wm$^{-2}$. $T_S$ is the surface temperature, $\Upsilon$ the amount of absorbing gas in g cm$^{-2}$, and $B_\nu(T)$ the Planck function. The isothermal broadband flux transmissivity $\varrho$, which is a function of temperature and optical depth, is defined as

$$ \varrho(\Upsilon, T) = \int_0^\infty \frac{\pi B_\nu(T) T_\nu(\Upsilon) d\nu}{\sigma_c T_k^4}. $$

A grey body approximation has been chosen to parameterize the longwave radiation, in which for every layer a single emissivity is used. Using the Stefan-Boltzmann law, Eqs. (2.12) and (2.13) may be rewritten in the forms

$$ F^\uparrow(\Upsilon) = \sigma_c T_S^4 \varrho(\Upsilon, T) + \int_0^\Upsilon \sigma_c T_k^4 (\Upsilon') d\varrho(\Upsilon - \Upsilon', T) $$

and

$$ F^\downarrow(\Upsilon) = \int_\Upsilon^\infty \sigma_c T_k^4 (\Upsilon') d\varrho(\Upsilon' - \Upsilon, T), $$

where $\sigma_c$ is the Stefan-Boltzmann constant and $T$ the absolute temperature. In defining these parameters, we assume that the plane-parallel atmosphere may be divided into thin isothermal layers. Moreover, the isothermal broadband flux emissivity $\epsilon$ is defined by

$$ \epsilon(\Upsilon, T) = 1 - \varrho(\Upsilon, T). $$

Using this formula, Eqs. (2.15) and (2.16) may be rewritten in the forms
\[ F^\uparrow(Y) = \sigma_c T^4 (1 - \epsilon(Y, T_S)) + \int_Y^0 \sigma_c T^4 (Y') \, d\epsilon(Y' - Y, T) \] (2.18)

and

\[ F^\downarrow(Y) = \int_Y^T \sigma_c T^4 (Y') \, d\epsilon(Y' - Y, T). \] (2.19)

For the emissivity the model uses the function

\[ \epsilon(Y, T) = \epsilon_c(Y_c, T) + \epsilon_{w1}(Y_{w1}, T) + \epsilon_{w2}(Y_{w2}, T). \] (2.20)

The emission functions are simplified versions (Cerni and Parish 1984) of the formula from Liou and Ou (1983). The emission function for CO₂ is

\[ \epsilon_c(Y_c, T) = \left(1.0 - |0.005T - 1.25|^{1.85}\right) \times \left(0.2167 + 0.005505 \log Y_c - 0.004189 \log^2 Y_c - 0.001328 \log^3 Y_c\right), \] (2.21)

with

\[ Y_c = \int_0^{Y_c} \frac{p}{p_0} \, dY, \] (2.22)

where \( p \) is the atmospheric pressure and \( p_0 \) the standard surface pressure, and with

\[ Y_c(z) = \int_0^z q_c \, dz, \] (2.23)

where the mixing ratio of the Carbon Dioxide \( q_c \) is defined as

\[ q_c = \frac{\rho_c}{\rho} \] (2.24)

with \( \rho_c \) for the density of the Carbon Dioxide.

The single emissivity that represents the 6.3 \( \mu \text{m} \) H₂O band, the 13 - 1000\( \mu \text{m} \) H₂O band, and the H₂O-CO₂ overlap is

\[ \epsilon_{w1}(Y_{w1}, T) = \left(0.7065 + 0.1431 \log Y_{w1} - 0.02228 \log^2 Y_{w1} - 0.004653 \log^3 Y_{w1}\right) \times \left(3.224 \log^2 T - 17.242 \log T + 23.62\right)^{1+0.1|\log Y_{w1}+1|}, \] (2.25)

with

\[ Y_{w1} = \int_0^{Y_w} \frac{p}{p_0} \, dY, \] (2.26)
and

\[ \Upsilon_w(z) = \int_0^z q_w \rho dz, \quad (2.27) \]

where the mixing ratio of the water vapor \( q_w \) is

\[ q_w = \frac{\rho_w}{\rho}, \quad (2.28) \]

with \( \rho_w \) for the density of the water vapor.

The emission function that represents the H$_2$O continuum is

\[ \epsilon_{w2}(\Upsilon_{w2}, T) = (0.8132 \log T - 1.741) \]

\[ \times \exp \left[ -\left( \frac{\log \Upsilon_{w2} - 8.892 \log T + 20.85}{1.597 \log T - 2.96} \right)^2 \right], \quad (2.29) \]

with

\[ \Upsilon_{w2} = \int_0^{\Upsilon_w} \frac{e}{e_0} d\Upsilon, \quad (2.30) \]

where \( e \) is the vapor pressure and \( e_0 \) the saturation vapor pressure at 296 K. The total emissivity \( \epsilon = \epsilon_e + \epsilon_{w1} + \epsilon_{w2} \).

The water vapor pressure is calculated according to Buck (1981),

\[ \epsilon_{0w} = \left[ 1.00072 + p \left( 3.20 \cdot 10^{-6} + 5.9 \cdot 10^{-10}(T - 273.15)^2 \right) \right] \]

\[ \times 6.112 \exp \left( \frac{18.729 - T - 273.15}{227.3} \right) \frac{T - 273.15}{T - 15.28}, \quad (2.31) \]

### 2.2.2 Shortwave radiation

The shortwave radiation is the main energy source at the surface. The ETH Expedition contributed to the development of a new parameterization, which has been tested with the measurements performed during two field seasons (Konzelmann et al., 1994). This parameterization, is used to calculate the shortwave net radiation at the surface,

\[ S = 0.84 S_0 \exp(-0.027 \tau_L m_r) \frac{1 - \alpha_s}{1 - \alpha_s \alpha_a}, \quad (2.32) \]

where \( \alpha_a \) is the albedo of the atmosphere, \( \alpha_s \) is the surface albedo, and \( \tau_L \) is the Linke turbidity factor. The relative optical air mass \( m_r \) is taken from Kasten (1966); it has been parameterized by Konzelmann et al. (1994) as
\[ m_r = \left[ \cos Z + \frac{0.15}{(93.885 - Z)^{1.253}} \right]^{-1}. \] (2.33)

The albedo becomes very important for the summer simulation. However there is no standard parameterization available. Over the ice sheet the albedo varies widely in space and time. Results from satellite measurements show a heterogeneous pattern in July (Haefliger 1995). The complexity of this pattern does not allow a simple parameterization of it.

In earlier researches, different attempts have been made to give the albedo as a function of the age of the snow (Hansen et al., 1983), the temperature, and the zenith angle (Kiehl et al., 1987). Further, there have been attempts to determine the albedo in a more sophisticated way as a function of snow grain size, snow depth, zenith angle, and snow cover pollutants (Marshall and Oglesby, 1994).

The parameterization from Segal et al. (1991) is presented as an example because it has been used in recently developed models. It only takes into consideration the effect of the zenith angle \( Z \). The albedo \( \alpha_S \) is

\[ \alpha_S = 0.8 + 0.32 \frac{1}{b} \left( \frac{b + 1}{1 + 2b \cos Z} - 1 \right). \] (2.34)

Gallée et al. (1996) propose a value of 2 for the constant \( b \). However, none of these parameterizations gave a satisfactory result. Thus, a new parameterization has been introduced. It is dependent of the latitude \( \gamma \) and on the height \( z \) above sea level:

\[ \alpha_S = \frac{1}{2} \left( 1.2 + 7.5 \times 10^{-5} \times z + 7 \left( \frac{1.1}{1 + 0.2 \cos \gamma} - 1 \right) \right). \] (2.35)

For the solar radiation at the top of the atmosphere the following formula is used (Iqbal 1983):

\[ S_0 = \frac{1}{r_E^2} S_\odot \cos Z, \] (2.36)

where the solar constant \( S_\odot = 1368 \, \text{Wm}^{-2} \), \( Z \) is the zenith angle and \( r_E \) is the normalized distance between earth and sun, and

\[ \frac{1}{r_E} = 1.00011 + 0.034221 \cos \Gamma + 1.28 \times 10^{-3} \sin \Gamma + 7.19 \times 10^{-4} \cos 2\Gamma + 7.7 \times 10^{-5} \sin 2\Gamma \] (2.37)

(Spencer 1971), where \( \Gamma \) is the day angle.

Above the surface, the solar heating of the air due the absorption of shortwave radiation becomes (in K s\(^{-1}\))

\[ \left( \frac{\partial T}{\partial t} \right)_{SW} = S \left( \frac{q}{c_p} \right) \left( \frac{p}{p_0} \right) \left[ \frac{\alpha (Y_{sw}(z, \infty))}{\cos Z} + 1.67 \alpha_s \frac{\alpha (Y_{sw})}{\cos Z} \right] + 1.7 \times 10^{-6} (\cos Z)^{0.3}, \] (2.38)
(Savijärvi 1990), where \( q \) stands for the absolute humidity; the function for the water vapor absorption in the reflected beam is

\[
\varpi(\Upsilon_w) = \begin{cases} 
0.029 \Upsilon_w^{0.81} & \Upsilon_w \geq 5 \cdot 10^{-4} m \\
0.050 \Upsilon_w^{0.63} & \Upsilon_w < 5 \cdot 10^{-4} m 
\end{cases}
\]  

where \( \Upsilon_w \) is the amount of water vapor.

### 2.2.3 Surface energy budget and surface temperature

The surface temperature is predicted from the force-restore slab model of Cerni and Parish (1984) and of Bhumralkar (1975):

\[
\frac{\partial T_S}{\partial t} = \begin{cases} 
\frac{1}{K_r} \left( F^1 - F^+ + S + Q_H + Q_{LE} \right) - \frac{K_r}{\tau} (T_S - T_m) + \frac{1}{K_g} K_L & : T_S \leq 0 \\
0 & : T_S > 0.
\end{cases}
\]  

The sensible heat flux is

\[
Q_H = \rho c_p F_{T_v},
\]  

with \( F_{T_v} \) for the surface layer; the latent heat flux is

\[
Q_{LE} = \rho c_p F_{Q_v},
\]  

and

\[
K_g = 0.95 \left( \frac{\lambda c_p}{2 \omega} \right)^{\frac{1}{2}}.
\]  

\( K_r \) has a value of 7.4, \( \lambda \) is the thermal conductivity, \( c_p \) the heat capacity of the snow, \( \omega \) is angular velocity of the earth's rotation, and \( \tau \) the length of the day. \( T_m \) is the ice temperature at a depth of 10 m, which is assumed to be constant over 24 hours. The model is not able to simulate the melting and freezing processes. To compensate this processes a correction factor \( K_L \) has been introduced which is assumed to be linear dependent from the height above sea level. For the summer simulation

\[
K_L = 45 - 0.0136 z,
\]  

and for the winter simulation

\[
K_L = 0.
\]
2.2.4 Planetary boundary layer formulation

Surface layer

In the model, the height of the surface layer corresponds to the first σ-level, which is about 11 m. However, Forrer and Rotach (1997) have shown that on the ETH/CU Camp the surface layer appears to be less than 10 m. Within the surface layer the friction terms \( F_{Uv} \) and \( F_{Vv} \), representing the frictional force, are treated as follows (Waight 1987):

\[
F_{Uv} = k_f V p^* u, \tag{2.46}
\]

\[
F_{Vv} = k_f V p^* v, \tag{2.47}
\]

where \( k_f \) is a friction coefficient with a value of \( 1.667 \times 10^{-5} \) and \( V \) the wind speed, and

\[
F_{Tv} = \frac{w' T}{\rho c_\rho} = \frac{Q_H}{\rho c_\rho} = -\frac{u_* k (T_1 - T_s)}{\ln \left( \frac{z}{z_{0H}} \right) - \Psi_H \left( \frac{z}{z_{0H}} \right)}, \tag{2.48}
\]

where \( T_1 \) is the temperature at the first σ-level, \( T_s \) the surface temperature, \( c_\rho \) the specific heat, \( \rho \) the density of the air, and \( z_{0H} \) the surface roughness length for the heat.

The friction velocity \( u_* \) is calculated from the logarithmic wind profile using the height \( z \) and the wind velocity \( V \), both from the first σ-level (Beljaars and Holtslag 1991):

\[
u_* = \frac{k V}{\ln \left( \frac{z}{z_M} \right) - \Psi_M \left( \frac{z}{z_M} \right)}, \tag{2.49}
\]

where \( k=0.4 \) is the von Karman constant, \( z_{0M} \) the roughness length for momentum, and \( L \) the Obukhov length scale given by

\[
L = \frac{z \phi_H}{Ri \phi_M^2}. \tag{2.50}
\]

Here \( \phi_H \) is the non-dimensional temperature gradient and \( \phi_M \) the non-dimensional velocity gradient. For computational reasons the gradient Richardson number is replaced by the bulk Richardson number:

\[
Ri_b = \frac{2g (\theta_1 - \theta_s) z}{(\theta_1 + \theta_s) V^2}. \tag{2.51}
\]

In the model, for the parameterization of \( \Psi_M \) and \( \Psi_H \) the formulas from Panofsky (1963) are taken for momentum

\[
\Psi_M (\zeta) = \int_0^\zeta [1 - \phi_M (\zeta)] \frac{d\zeta}{\zeta} \tag{2.52}
\]

and for heat
\[ \Psi_H (\zeta) = \int_{\zeta_0}^{\zeta} [1 - \phi_H (\zeta)] \frac{d\zeta}{\zeta}, \]  
(2.53)

where we define

\[ \zeta = \frac{z}{L} \quad \text{and} \quad \zeta_0 = \frac{z_0}{L}. \]  
(2.54)

For the unstable case, the dimensionless stability function for momentum is (Paulson 1970)

\[ \Psi_M = 2 \ln \left( \frac{1 + A}{2} \right) + \ln \left( \frac{1 + A^2}{2} \right) - 2 \arctan A + \frac{\pi}{2} \]  
(2.55)

and for heat

\[ \Psi_H = 2 \ln \left( \frac{1 + A^2}{2} \right), \]  
(2.56)

with

\[ A = \left( 1 - 16 \frac{z}{L} \right)^{\frac{1}{4}}. \]  
(2.57)

For the stable case, the dimensionless stability function for momentum is (Beljaars and Holtslag 1991)

\[ \Psi_M (\zeta) = -a \frac{z}{L} - b \left( \frac{z}{L} - \frac{c}{d} \right) e^{-d \frac{\zeta}{d}} - \frac{bc}{d}, \]  
(2.58)

with \( a = 0.7, b = 0.75, c = 5, \) and \( d = 0.35. \) The dimensionless stability function for heat is

\[ \Psi_H (\zeta) = - \left( 1 + \frac{2 az}{3 L} \right)^{\frac{3}{2}} - b \left( \frac{z}{L} - \frac{c}{d} \right) e^{-d \frac{\zeta}{d}} - \frac{bc}{d} + 1, \]  
(2.59)

with \( b = 0.667, c = 5 \) and \( d = 0.35. \) In addition, to compute the surface temperature an equation for the evaporation/sublimation is needed. The term is \( \overline{w'q'} \) and it is calculated in an analogous way as \( \overline{w'T'} \), that is

\[ \overline{w'q'} = \frac{Q_{LE}}{\overline{\ell_o \rho}} = \frac{-u^* k (q_1 - q_s)}{\ln \left( \frac{z}{z_{0H}} \right) - \Psi_H (\frac{\zeta}{L})}, \]  
(2.60)

where \( q_1 \) is the specific humidity at the first \( \sigma \)-level, \( q_s \) the specific humidity at the surface, \( \overline{\ell_o} \) the latent heat of vaporization, and \( z_{0H} \) the surface roughness length for heat.
Planetary boundary layer

Above the surface layer the flux for momentum, temperature, and humidity is treated as follows:

\[
F_{Uu} = -\frac{\partial p_u w' v'}{\partial z} = \frac{\partial}{\partial z} \left( K \frac{\partial p_u}{\partial z} \right),
\]

\[
F_{Vv} = -\frac{\partial p_w w' v'}{\partial z} = \frac{\partial}{\partial z} \left( K \frac{\partial p_w}{\partial z} \right),
\]

\[
F_{Tt} = -\frac{\partial w' \theta'}{\partial z} = \frac{\partial}{\partial z} \left( K \frac{\partial \theta}{\partial z} \right),
\]

\[
F_{Qq} = -\frac{\partial w' q'}{\partial z} = \frac{\partial}{\partial z} \left( K \frac{\partial q}{\partial z} \right).
\]

Therefore, a profile for the eddy diffusivity \( K \) (assumed to be the same for heat, moisture, and momentum) is needed. The diagnostic equation for the height of the stable boundary layer (Brost and Wyngaard, 1978) is

\[
h = d \left( \frac{u_* L}{f} \right)^{-\frac{1}{2}},
\]

with the constant \( d = 0.4 \). Now \( K \) is calculated from the profile equation of Brost and Wyngaard (1978) as

\[
K = \frac{k u^* z \left( 1 - \frac{z}{h} \right)^{-\frac{2}{3}}}{1 + 4.7 \frac{z}{L}}.
\]

2.2.5 Horizontal diffusion

Horizontal diffusion for momentum and temperature is necessary to ensure computational stability. Second order diffusion terms are introduced on the right-hand side of Eqs. (2.6) to (2.8) to parameterize sub-grid scale diffusion along the \( \sigma \)-surfaces. They have the form

\[
F_{Uu} = \nu_{h_m} \frac{\partial^2 p_u}{\partial x^2} + \nu_{h_m} \frac{\partial^2 p_u}{\partial y^2},
\]

\[
F_{Vv} = \nu_{h_m} \frac{\partial^2 p_w}{\partial x^2} + \nu_{h_m} \frac{\partial^2 p_w}{\partial y^2},
\]

\[
F_{Tt} = \nu_{h_T} \frac{\partial^2 \theta}{\partial x^2} + \nu_{h_T} \frac{\partial^2 \theta}{\partial y^2},
\]

where \( \nu_{h_m} \) is an horizontal diffusion coefficient for momentum and \( \nu_{h_T} \) for temperature. They are given by Waight (1987) as
\begin{equation}
\nu_{hm} = \bar{c}_v + c'_v V, \quad \nu_{hT} = \bar{c}_t + c'_t V,
\end{equation}

with

\begin{equation}
\bar{c}_v = 0.075 \Delta x,
\end{equation}

\begin{equation}
c'_v = 0.375 \Delta x,
\end{equation}

\begin{equation}
\bar{c}_t = 0.015 \Delta x,
\end{equation}

\begin{equation}
c'_t = 0.075 \Delta x,
\end{equation}

where \( \Delta x \) is the model grid spacing in meters.

\section*{2.3 Boundary conditions}

\subsection*{2.3.1 Lateral boundary conditions}

The treatment of lateral boundary conditions is important for the stability of the model. The mesoscale model of Anthes and Warner (1978) has shown that the mean acceleration of the flow is very sensitive to low temperature and pressure differences across the domain. For this reason \( p_* \) is specified on the lateral boundaries. The temperature on the boundaries is allowed to change only in the case of radiative and turbulent flux divergence. The velocity components are extrapolated to the boundaries from the interior of the domain. The vertical velocity \( \dot{\sigma} \) is set to zero on all lateral boundaries. This approximation is allowed because all boundaries of the model are above the homogeneous ocean surface.

To avoid numerical reflection on the boundaries, the so-called porous sponge method of Perkey and Kreitzberg (1976) is used. The tendency in the prognostic equations is multiplied by a weighting factor which decreases from a value of one in the interior of the domain to nearly zero on the lateral boundaries.

For the winter simulation, an 80 km large area on each boundary is influenced by this factor. For the summer simulation this area corresponds to 160 km. The entire model surface is 1640 km x 2800 km.

\subsection*{2.3.2 Top of the model}

The upper boundary of the model has been chosen high enough to exclude the influence of the atmosphere above the model to the planetary boundary layer. Nevertheless, some conditions have to be specified:
\[ S = S_0, \]
\[ \dot{\sigma} = 0, \]
\[ F^\dagger + F^\dagger = 0, \]
\[ \frac{\partial \theta}{\partial \sigma} = \frac{\partial q}{\partial \sigma} = \frac{\partial u}{\partial \sigma} = \frac{\partial v}{\partial \sigma} = 0. \]

For both runs presented in this work the top \( \sigma \)-level is at 250 mbar.
3 Field observations

For a comparison between model output and field observations, measurements from four automatic weather stations were used. The location of the stations is shown in Fig. 3.1.

![Map of Greenland showing the position of the automatic weather stations](image)

Figure 3.1: A map of Greenland showing the position of the automatic weather stations i.e. ETH/CU Camp, Humboldt, Tunu, and GISP2.

At the ETH/CU Camp the measurements started in summer 1990 and are still running in 1998. In 1995, the University of Colorado, Boulder and the Cooperative Institute for Research in Environmental Sciences (CIRES) started research with an automatic weather station network on the ice sheet. This network includes at the moment 14 automatic weather stations. Results from two of these stations were used in this work. The first
station, Humboldt, is situated north of Thule. The other, Tuni, is situated on the north-east slope. In addition to this network in June 1989 a unit was installed close to the summit by the Space Science and Engineering Center of the University of Wisconsin in Madison, Wisconsin. The name of the station is GISP2.

3.1 ETH/CU Camp

The ETH/CU Camp is situated in the Paqitsoq area on the western slope of Greenland near the average equilibrium line (69° 34' 25" North, 49° 17 '44" West, 1155 m a.s.l.). From June 1 to September 5, 1990 and from May 1 to August 30, 1991 the Institute of Geography of the ETH in Zurich conducted a large scientific program. During these two summers all components of energy and mass balance were measured. After the second field season the measurements were carried out by an automatic weather station (AWS). Since 1993 the AWS has been maintained by CIRES.

The following sections describe the measurements and discuss some of the results which are relevant to this work.

3.1.1 Wind measurements

The wind speed sensors and wind vanes were fixed on two towers with a height of ten and thirty meters. On the top of the ten-meter tower an anemometer and a wind vane were fixed. The thirty-meter tower was equipped with cup anemometers on eight levels and with wind vanes on three levels. The nominal heights of the anemometers were 0.5 m, 1 m, 2 m, 5 m, 10 m, 15 m, 22.5 m, and 30 m and that of the wind vanes 1 m, 15 m, and 30 m (Ohmura et al., 1991).

During the two field seasons the wind speed was integrated over ten minutes, and these values were stored. The wind direction was measured every minute, and the ten-minute mean values were stored.

After the summer season 1991, when the automatic measurements started, wind speed and wind direction were determined from the ten-meter tower. The values were taken every ten minutes and mean values over three hours were stored.

Periods with data are for:

- wind speed at 10 m:
  10 min values: 1.6.90 - 5.9.90 / 1.5.91 - 31.8.91
  3 h values: 1.9.91 - 8.6.93 / 4.9.94 - 23.5.95 / 19.6.95 - 7.6.96

- wind direction at 10 m:
  10 min values: 1.6.90 - 5.9.90 / 1.5.91 - 30.8.91
  3 h values: 19.6.92 - 14.11.92 / 7.2.93 - 16.4.93 / 24.6.93 - 23.5.95 / 19.6.95 - 7.6.96

- wind speed and direction at 30 m:
  10 min values: 15.6.90 - 29.8.90 / 3.6.91 - 19.8.91.
The monthly mean values for the wind speed at 10 m are presented in Table 3.1. Only months without or with few missing values were considered. The measurements show that the values for wind speed are larger in winter than in summer. From June 1990 to May 1996 the lowest value was recorded in July 1995 with 4.9 m s\(^{-1}\) and the highest value in April 1995 with 11.4 m s\(^{-1}\). The wind speed difference between the same month but in different years can be as large as 4.4 m s\(^{-1}\) for April. In some years the monthly change is steady, like in fall 1991 or spring 1993; in other years the mean wind speed changes are not so smooth, like in fall 1992 or spring 1995.

Table 3.1: Monthly mean wind speed at 10 m above the surface on the ETH/CU Camp. Measurements from June 1990 to May 1996; unit: m s\(^{-1}\). *: months with no or only few valid data.

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<tbody>
<tr>
<td>January</td>
<td>10.4</td>
<td>9.9</td>
<td>*</td>
<td>10.1</td>
<td>8.8</td>
<td></td>
<td></td>
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<tr>
<td>February</td>
<td>9.0</td>
<td>9.9</td>
<td>*</td>
<td>9.6</td>
<td>7.8</td>
<td></td>
<td></td>
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<tr>
<td>March</td>
<td>8.3</td>
<td>8.8</td>
<td>*</td>
<td>9.1</td>
<td>6.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>April</td>
<td>10.2</td>
<td>8.5</td>
<td>*</td>
<td>11.4</td>
<td>7.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>May</td>
<td>6.7</td>
<td>8.0</td>
<td>*</td>
<td>7.4</td>
<td>6.1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>June</td>
<td>8.4</td>
<td>6.9</td>
<td>7.1</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td></td>
</tr>
<tr>
<td>July</td>
<td>7.1</td>
<td>7.1</td>
<td>6.8</td>
<td>*</td>
<td>*</td>
<td>4.9</td>
<td></td>
</tr>
<tr>
<td>August</td>
<td>7.9</td>
<td>7.3</td>
<td>7.4</td>
<td>*</td>
<td>*</td>
<td>5.3</td>
<td></td>
</tr>
<tr>
<td>September</td>
<td>7.4</td>
<td>10.6</td>
<td>*</td>
<td>7.5</td>
<td>8.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>October</td>
<td>8.2</td>
<td>8.1</td>
<td>*</td>
<td>7.5</td>
<td>8.6</td>
<td></td>
<td></td>
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<tr>
<td>November</td>
<td>9.1</td>
<td>11.2</td>
<td>*</td>
<td>10.6</td>
<td>8.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>December</td>
<td>10.8</td>
<td>11.0</td>
<td>*</td>
<td>10.1</td>
<td>7.3</td>
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</table>

The mean daily wind speed cycle for each month is shown in Fig. 3.2. In January, February, November and December the wind speed values do not show a daily cycle. The results are presented for local standard time. In March, with increasing solar radiation, a daily cycle starts to build up and reaches a maximum amplitude of 1.4 m s\(^{-1}\) in April. From May to October this amplitude decreases and disappears in November, when the solar radiation becomes very weak. The difference between the individual years is very large in April and decreases to almost zero in October. Daily cycles of wind speed show a maximum early in the morning and a minimum in the evening.

The values of the three-hourly mean wind direction for the year 1995 is presented in Fig. 3.3. The year 1995 has been chosen because it contains the least number of missing data, namely 32 from 2920 possible values. The upper graph shows the distribution of all data split into segments of ten degrees. The main wind direction is from south-west, with 90.2 % of all data between 90 and 180 degrees (360 degrees correspond to a wind from north). The segment between 120 and 130 degrees occurs most frequently with 492 values (17.0 %).

The data from summer (April to September) and winter (October to March) are presented
in the two lower graphs in Fig. 3.3 (left: summer; right: winter). In summer the wind blows more frequently from the south than during winter. The downslope vector is estimated to be 90 (± 15) degrees (Ohmura et al., 1991). This means that the wind turns about 35 degrees to the right from the downslope direction in summer and 15 degrees in winter. The comparison with the downslope direction has a large uncertainty because it includes an estimate on a scale of 1 km (80 (± 5) degrees) and an estimate on a scale of 10 km (100 (± 5) degrees).

Table 3.2: Monthly directional constancy of the wind at the ETH/CU Camp. *: months with no or only few valid data.

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<td>*</td>
<td>0.96</td>
<td>0.92</td>
<td></td>
<td></td>
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<td>February</td>
<td></td>
<td>*</td>
<td>*</td>
<td>0.96</td>
<td>0.88</td>
<td></td>
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<td>*</td>
<td>0.93</td>
<td>0.86</td>
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<td>*</td>
<td>0.94</td>
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<td>*</td>
<td>*</td>
<td>0.93</td>
<td>0.83</td>
<td></td>
<td></td>
</tr>
<tr>
<td>June</td>
<td>0.95</td>
<td>0.92</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td></td>
<td></td>
</tr>
<tr>
<td>July</td>
<td>0.95</td>
<td>0.92</td>
<td>0.89</td>
<td>*</td>
<td>0.91</td>
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<tr>
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<td>0.92</td>
<td>0.85</td>
<td>0.88</td>
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<td></td>
<td>0.92</td>
<td>*</td>
<td>0.86</td>
<td>0.90</td>
<td></td>
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<tr>
<td>October</td>
<td></td>
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<td>*</td>
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<td>0.89</td>
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</tr>
<tr>
<td>November</td>
<td></td>
<td>*</td>
<td>*</td>
<td>0.95</td>
<td>0.86</td>
<td></td>
<td></td>
</tr>
<tr>
<td>December</td>
<td></td>
<td>*</td>
<td>*</td>
<td>0.92</td>
<td>0.85</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The directional constancy is defined as the ratio of the vector mean of the wind speed divided by the mean wind speed (Schwerdtfeger 1984). The values are presented in Table 3.2. In katabatic wind regimes this constancy is close to one. The values for the ETH/CU Camp are comparable to the values observed in other katabatic wind regimes as observed on the Antarctic ice sheet. Schwerdtfeger (1984) determines values of 0.97 for Cape Denison, 0.90 for Mirny, and 0.85 for Molo 0.85. All three stations are close to the coast at the bottom of the slope of the continental ice sheet. Bromwich (1989) presents values between 0.94 and 0.99 for the station at the southern end of Inexpressible Island. Parish (1982) reports 0.92 for Pionerskaya, 0.91 for Charcot, and 0.86 for Byrd. All these stations are on the slope of the Antarctic ice sheet. Wendler and Kodama (1984) present monthly mean values for six stations in Antarctica situated between the center of the continent and the coast. The directional constancy for the annual mean ranges between 0.51 at 3280 m a.s.l (Dome C) and 0.94 at 14 m a.s.l (Port Martin). It indicates smaller monthly variability for stations close to the margin of the ice sheet than for stations located towards the center of the continent. The station D57, situated at 2103 m on the transition from the gentle to the steep slope, shows similar values to the ETH/CU Camp.
Figure 3.2: Mean daily wind speed cycle for each month taken from a ten-meter level at the ETH/CU Camp. The solid lines represent the mean values over all available data. The dotted lines represent the individual years.
Figure 3.3: The wind-rose at the top shows the distribution of 2888 three-hourly values for 1995 at the ETH/CU Camp. The lower left wind-rose shows the distribution of the values from April to September 1994; the lower right wind-rose shows the values from January to March and from October to December 1995. (One segment represents all values within ten degrees.)
3.1.2 Temperature measurements

Temperature measurements were obtained from a thermistor at a nominal height of two meters on the ten-meter mast and on all eight levels on the 30-meter tower. In 1990 and after August 1991, the instruments were not ventilated, however they were shielded from radiation. Between the period of May and August 1991 the instruments were ventilated. During the two main field seasons readings were taken every 15 seconds and averaged for half-hourly means. After August 91 only three-hourly means were recorded. Data is available for

- temperature on the 10 m mast:
  - 10 min values: 1.6.90 - 5.9.90 / 1.5.91 - 31.8.91
  - 3 h values: 1.9.91 - 8.6.93 / 31.7.93 - 23.5.95 / 19.6.95 - 7.6.96

- temperature on the 30 m tower:
  - 10 min values: 15.6.90 - 29.8.90 / 3.6.91 - 19.8.91.

The monthly mean temperature is presented in Table 3.3. The difference between the individual years is rather large. The winter 1995/96 was exceptionally warm compared to earlier years. February 1992 and March 1993 were the coldest months with a temperature below -30 degrees. In 1990, 1991 and 1995 July was the only month with a mean temperature above zero degrees, while in 1992 and 1994 the monthly mean temperature was never above zero.

The mean daily temperature cycle for each month is summarized in Fig. 3.5. The seasonal variation of the amplitude is of the “Frame” type. This indicates that the year’s maximum diurnal temperature amplitude occurs in spring. Often there is another peak in late summer which is less prominent (Simpson 1919; Ohmura 1984). The maximum amplitude at the ETH/CU Camp is in May.

The ice temperature in ten different depths was measured once a day (Niederbäumer 1994). The nominal depths of the sensors were 0.25 m, 0.75 m, 1 m, 2 m, 3 m, 4 m, 5 m, 6 m, 8 m, and 10 m. In Fig. 3.4 the temperature profile for the months January, April, July and October are shown. At a ten meter depth the annual amplitude is less than one degree and becomes 5.7 degrees on the surface. In January the gradient of the top two meters is very steep and disappears in the lower layers. From January to April energy vanishes through the surface and the entire layer gets colder. From April to July the sun heats up the first four meters; this process continues through October, when all ten meters are heated. In October the very top layer starts losing energy and an inversion begins to build up. During winter this inversion disappears and the surface gets colder compared to the deeper layers.
Table 3.3: Monthly mean air temperature at 2 m above the surface at the ETH/CU Camp. Measurements from June 1990 to May 1996; unit: °C. Months with no or only few valid data are marked by an asterix (*).

<table>
<thead>
<tr>
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<td>-17.4</td>
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</tr>
<tr>
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<td>*</td>
<td>-4.9</td>
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</tr>
<tr>
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<td>*</td>
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<td>0.9</td>
<td></td>
</tr>
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<td>-1.4</td>
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<td>-8.9</td>
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<td>-14.8</td>
<td>-16.5</td>
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</tr>
<tr>
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<td>-22.7</td>
<td>-17.8</td>
<td>-10.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>December</td>
<td>-20.5</td>
<td>-24.6</td>
<td>-25.5</td>
<td>-27.2</td>
<td>-18.4</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 3.4: Mean monthly ice temperature profile of the first ten meters at the ETH/CU Camp. From left to right: January 1992 - 95, April 1992 - 95, July 1990 - 92/94, October 1991 - 94.
Figure 3.5: Mean daily temperature cycle for each month from a ten-meter level at the ETH/CU Camp. The solid lines represent the mean value over all available data. The dotted lines represent the individual years.
### 3.1.3 Energy flux

The radiation flux density was measured for the periods June 6 to September 2, 1990 and May 9 to August 29, 1991. Measurements were performed for direct solar radiation, diffuse sky radiation, global radiation, shortwave reflected radiation, longwave incoming radiation, longwave outgoing radiation, and net radiation. The reflected shortwave radiation for the albedo determination was measured at 2 m and, on the tower, at 27 m above ground. Detailed cloud observations were usually made seven times a day (0, 3, 6, 12, 15, 18, and 21 UTC). A detailed description of the radiation condition at the ETH/CU Camp is given in Konzelmann (1994). The monthly means are presented in Table 3.4.

#### Table 3.4: Monthly mean values for the radiation components in Wm$^{-2}$ at the ETH/CU Camp for 1990 and 1991. The period of measurements was from June 6 to September 9, 1990 and from May 9 to August 29, 1991. The last column (season) includes the values from June 6 to August 29 for both years. The values in brackets are the number of days with measurements. The periods of relative calibrations were not taken into consideration (Ohmura et al., 1992).

<table>
<thead>
<tr>
<th>Components</th>
<th>May</th>
<th>June</th>
<th>July</th>
<th>August</th>
<th>Season</th>
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<tr>
<td></td>
<td>91</td>
<td>90</td>
<td>91</td>
<td>90</td>
<td>91</td>
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<tr>
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<td>378</td>
<td>334</td>
<td>307</td>
<td>320</td>
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<tr>
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<td>285</td>
<td>184</td>
<td>195</td>
<td>216</td>
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<tr>
<td>Diffuse sky radiation</td>
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<td>93</td>
<td>150</td>
<td>112</td>
<td>104</td>
</tr>
<tr>
<td>Shortwave reflected radiation</td>
<td>-277</td>
<td>-278</td>
<td>-255</td>
<td>-210</td>
<td>-236</td>
</tr>
<tr>
<td>Albedo at ground level</td>
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<td>0.74</td>
<td>0.76</td>
<td>0.68</td>
<td>0.74</td>
</tr>
<tr>
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<td>233</td>
<td>260</td>
<td>255</td>
<td>256</td>
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<tr>
<td>Longwave outgoing radiation</td>
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<td>-311</td>
<td>-307</td>
<td>-312</td>
<td>-310</td>
</tr>
<tr>
<td>Shortwave net radiation</td>
<td>46</td>
<td>99</td>
<td>78</td>
<td>96</td>
<td>83</td>
</tr>
<tr>
<td>Longwave net radiation</td>
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<td>-78</td>
<td>-47</td>
<td>-57</td>
<td>-54</td>
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<tr>
<td>Net radiation</td>
<td>-1</td>
<td>22</td>
<td>32</td>
<td>40</td>
<td>30</td>
</tr>
</tbody>
</table>

The monthly mean values for the global radiation in 1990 are largest in June and decrease in July and August. In 1991 the values increase from May to June and decrease in July and August. The seasonal global radiation was 4 % smaller in 1991 than in 1990. It can thus be concluded that in 1991 the lack of direct solar radiation was almost compensated by diffuse sky radiation. The monthly mean albedo can change by more than 10 % within one year. Small fluctuations were found for the longwave radiation.
3.2 Humboldt and Tunu

In summer 1995 CIRES installed an automatic weather station named "Humboldt" on the north west slope in northern Greenland at 78° 35' North and 57° 13' West. In summer 1996 the station "Tunu" was installed on the north east slope at 78° 01' North and 33° 59' West. Both stations are situated at 2000 m above sea level. The measurements were taken every 10 minutes and hourly means were stored. The air temperature, wind speed, and humidity were measured 2.3 m and 4.1 m above ground. Other parameters measured were pressure, snow temperature, and wind direction. Energy terms sampled were net radiation, shortwave incoming and reflected shortwave radiation. The height above ground corresponds to the values at the beginning of the measurements on June 22, 1995 for Humboldt and May 18, 1996 for Tunu.

The monthly mean average for Humboldt is presented in Table 3.5. For temperature and wind speed the values are taken from the sensors at 4.1 m. The temperature is between 5 and 15 degrees lower than at the ETH/CU Camp. The monthly difference is similar in both stations. The values for the wind speed are similar to those at the ETH/CU Camp. The largest difference between the two stations has been detected for October 1995 with 2 m s^{-1}, however, most of the time the difference is smaller than 1 m s^{-1}. Particularly in winter the directional constancy is very high. The monthly means of the net radiation are negative except for March 1996; the high albedo varies little over the seasons. The negative values for the net radiation in the transition seasons and summer season is determined by the high albedo and smaller values for the outgoing longwave radiation compared to the ETH/CU-Camp. Humboldt represents a point on the ice sheet where the katabatic wind blows most of the time and the temperature is below zero.

In Fig. 3.6 the wind-rose for Humboldt is shown for the year 1996. The annual distribution of all hourly values (8760 values in total) shows a pattern which is characteristic for katabatic wind regime; 26 % of the values are between 200 and 220 degrees, 82 % between 160 and 250 degrees. During the six summer months the wind direction is more likely to vary between south and east, but there are still 78 % of cases with values between 160 and 250 degrees. During the winter months there are even more values between 200 and 220 degrees than in the annual average, namely 32 %. The downslope vector is estimated to be 150 (± 15) degrees. This means the wind turns about 60 degrees to the right from the downslope direction.

The monthly mean values from the automatic weather station in Tunu are presented in Table 3.6. The period with available data goes from June 1996 to September 1997. The results correspond to those of the other stations and present values characteristic for katabatic regimes. The wind speed is between 4.4 m s^{-1} in July 1997 and 8.9 m s^{-1} in January 1997. It shows a high directional constancy with values of 0.71 in July 1997 and 0.95 in May 1997. The monthly mean temperature is always far below zero. The values for the radiation are similar to those found at Humboldt.

The wind-rose for Tunu is presented in Fig. 3.7. The main wind direction comes from the west with 47 % of the values between 260 and 290 degrees. The downslope vector is estimated to be 240 (± 15) degrees. This means that the wind turns about 35 degrees
Table 3.5: Values from the AWS at Humboldt: monthly mean air temperature (T), wind speed (WS), directional constancy (DC), incoming shortwave radiation (SWin), reflected shortwave radiation (Wout), net radiation (Net), Albedo (A). Measurements from July 1995 to September 1997 (CIRES data)

<table>
<thead>
<tr>
<th>Year</th>
<th>Month</th>
<th>T (°C)</th>
<th>WS (m s⁻¹)</th>
<th>DC</th>
<th>SWin (W/m²)</th>
<th>SWout (W/m²)</th>
<th>Net (W/m²)</th>
<th>A (%)</th>
</tr>
</thead>
<tbody>
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<td>0.87</td>
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<tr>
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<tr>
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<tr>
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<tr>
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<td>0</td>
</tr>
<tr>
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<td>0.93</td>
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<td>3</td>
<td>96</td>
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<tr>
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<td>0.87</td>
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<td>90</td>
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<td>190</td>
<td>-174</td>
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<td>0.92</td>
<td>80</td>
<td>-72</td>
<td>-25</td>
<td>90</td>
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</table>

to the right from the downslope direction. There is no significant difference between the summer and winter season.
Figure 3.6: The wind-rose on the top shows the distribution of all hourly values for 1996 at Humboldt. The lower left wind-rose shows the distribution of the values from April to September 1996; the lower right wind-rose shows the values for January to March and for October to December 1996. (One segment represents all values within ten degrees.)
Table 3.6: Values from the AWS at Tunu: monthly mean air temperature ($T$), wind speed ($WS$), directional constancy ($DC$), incoming shortwave radiation ($SW_{in}$), reflected shortwave radiation ($SW_{out}$), net radiation ($Net$), Albedo ($A$). Measurements from June 1996 to September 1997 (CIRES data).

<table>
<thead>
<tr>
<th>Month</th>
<th>$T$ (°C)</th>
<th>$WS$ (m s$^{-1}$)</th>
<th>$DC$ (%)</th>
<th>$SW_{in}$ (W/m$^2$)</th>
<th>$SW_{out}$ (W/m$^2$)</th>
<th>Net (W/m$^2$)</th>
<th>$A$ (%)</th>
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</thead>
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<td>-12</td>
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<tr>
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<td>-</td>
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<td>-9</td>
<td>-</td>
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<tr>
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<td>-157</td>
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<tr>
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<td>328</td>
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<td>-22</td>
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<tr>
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<td>391</td>
<td>-349</td>
<td>-15</td>
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<tr>
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<td>0.71</td>
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<td>-303</td>
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</tr>
<tr>
<td>August 97</td>
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<td>0.87</td>
<td>212</td>
<td>-191</td>
<td>-19</td>
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<tr>
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<td>0.85</td>
<td>74</td>
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<td>-15</td>
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</table>
Figure 3.7: The wind-rose on the top shows the distribution of hourly values for April 1996 to March 1997 at Tunu. The lower left wind-rose shows the distribution of the values from April to September 1996; the lower right wind-rose shows the values for January to March 1997 and for October to December 1996. (One segment represents all values within ten degrees.)
3.3 GISP2

Automatic weather stations are installed on the crest of Greenland (38.46 West, 72.58 North, 3205 m a.s.l.). These are supported by the ice coring activities of the United States and Western Europe. In June 1989 a unit was installed close to the American drill-hole by the Space Science and Engineering Center of the University of Wisconsin in Madison, Wisconsin. The station was called GISP2, after the drilling site. The AWS measures wind speed, wind direction, air temperature, relative humidity at a nominal height of 3.6 m, air pressure, and the vertical temperature difference between 0.5 and 3.6 m. Three-hourly mean values are available on the World Wide Web.

Table 3.7: Monthly mean air temperature at GISP2. Measurements from June 1989 to May 1995; unit: °C. Months with more than 50 % of missing values are indicated by an asterix (*). (Data from the University of Wisconsin)

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<tbody>
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<td>January</td>
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<td>-42.2</td>
<td>-40.6</td>
<td>-45.3</td>
<td>-44.4</td>
<td></td>
</tr>
<tr>
<td>February</td>
<td>-44.2</td>
<td>-36.8</td>
<td>*</td>
<td>-38.5</td>
<td>-43.0</td>
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<tr>
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<td>-38.9</td>
<td>*</td>
<td>-42.2</td>
<td>-42.6</td>
<td></td>
</tr>
<tr>
<td>April</td>
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<td>-37.0</td>
<td>*</td>
<td>-28.4</td>
<td>-29.4</td>
<td></td>
</tr>
<tr>
<td>May</td>
<td>-18.1</td>
<td>-19.9</td>
<td>*</td>
<td>-21.1</td>
<td>-21.8</td>
<td></td>
</tr>
<tr>
<td>June</td>
<td>-13.9</td>
<td>-13.6</td>
<td>-11.5</td>
<td>*</td>
<td>-17.4</td>
<td>-14.1</td>
</tr>
<tr>
<td>July</td>
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<td>-9.6</td>
<td>-14.8</td>
<td>-11.7</td>
<td>-11.5</td>
</tr>
<tr>
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<td>*</td>
<td>-19.8</td>
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<tr>
<td>September</td>
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<td>-24.7</td>
<td>-26.3</td>
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<tr>
<td>October</td>
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<td>-33.7</td>
<td>-31.9</td>
<td>-26.7</td>
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<td>-32.7</td>
<td>-29.7</td>
<td>-40.6</td>
<td>-35.7</td>
<td>-37.4</td>
<td></td>
</tr>
<tr>
<td>December</td>
<td>-36.5</td>
<td>-38.1</td>
<td>-39.6</td>
<td>-45.9</td>
<td>-42.8</td>
<td></td>
</tr>
</tbody>
</table>

In Table 3.7 the monthly mean air temperatures are displayed. All values are below zero degrees. Remarkable is the fact that for 1990 the mean temperature in November is higher than in October; for 1989 and 1994, though, both months have the same temperature. This anomaly for November was also observed at Humboldt and on the ETH/CU Camp in 1995. A further observation in this table is that the year to year temperature difference in winter is larger than in summer.

In Table 3.8 the monthly mean wind speed and the directional constancy are presented. The wind speed is lower than that registered by the other two stations. Usually, a high directional constancy corresponds to high wind speed. The directional constancy is much smaller than that recorded by the stations on the slope. Also, it is larger in winter than in summer. An exception is December 1992, with a very low directional constancy of 0.28. Another exception is February 1995, where the directional constancy is high while the wind speed is low. However, the data file with the hourly values shows a period of five days with no wind. A failure of the sensor can not be excluded.
Table 3.8: Monthly mean wind speed (WS) and directional constancy (DC) at GISP2. Measurements from June 1989 to May 1995; unit: m s\(^{-1}\). Months with more than 50% missing values are indicated by an asterix (*). (Data from the University of Wisconsin)

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<td>DC</td>
<td>WS</td>
<td>DC</td>
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<td>DC</td>
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<td>January</td>
<td></td>
<td>*</td>
<td>5.9</td>
<td>0.77</td>
<td>8.6</td>
<td>0.82</td>
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<tr>
<td>February</td>
<td></td>
<td>*</td>
<td>6.0</td>
<td>0.78</td>
<td></td>
<td></td>
</tr>
<tr>
<td>March</td>
<td></td>
<td>*</td>
<td>4.6</td>
<td>0.65</td>
<td></td>
<td></td>
</tr>
<tr>
<td>April</td>
<td></td>
<td>*</td>
<td></td>
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<td>*</td>
<td></td>
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<td></td>
<td>*</td>
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<td>June</td>
<td>3.3</td>
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<tr>
<td>July</td>
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<td>0.08</td>
<td></td>
<td></td>
<td>2.9</td>
<td>0.42</td>
</tr>
<tr>
<td>August</td>
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<td>0.55</td>
<td>3.7</td>
<td>0.46</td>
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<tr>
<td>September</td>
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<td>0.25</td>
<td>4.3</td>
<td>0.22</td>
<td>4.3</td>
<td>0.31</td>
</tr>
<tr>
<td>October</td>
<td>3.1</td>
<td>0.67</td>
<td>3.9</td>
<td>0.82</td>
<td>5.6</td>
<td>0.64</td>
</tr>
<tr>
<td>November</td>
<td>5.7</td>
<td>0.80</td>
<td>4.9</td>
<td>0.91</td>
<td>4.8</td>
<td>0.41</td>
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<tr>
<td>December</td>
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<td>*</td>
<td>5.4</td>
<td>0.28</td>
<td>8.6</td>
<td>0.80</td>
</tr>
</tbody>
</table>

In Fig. 3.8 the wind-rose for GISP2 is shown for the year 1994. The annual distribution of all three-hourly values shows a different pattern as the three other wind-roses presented earlier in this chapter. There is no dominant wind direction, like at the stations on the slope.
Figure 3.8: The wind-rose on the top shows the distribution of hourly values for 1994 at GISP2. The lower left wind-rose shows the distribution of the values from April to September 1994; the lower right wind-rose shows the values for January to March 1994 and for October to December 1994. (One segment represents all values within ten degrees).
4 Winter Simulation

4.1 Introduction

In this work two simulations will be described. The first represents the conditions during polar winter with no solar radiation north of the polar circle; the second represents the polar summer with maximum solar radiation. Most of the investigations with numerical simulation of katabatic wind have been done for the winter season in Antarctica. During this period the katabatic wind regime is fully developed and stationary over a long period of time.

There is a difference between these two ice sheets in respect to their degrees of latitude. The Antarctic ice sheet covers an area from about 70° to 90°S. The Greenland ice sheet is between 60° and 83°N. This leads to much colder temperatures and a lower moisture content of the air for Antarctica.

The Greenland ice sheet extends from 60°N to 83°N and from 12°W to 73°W. The ice sheet is presently 1.75×10^6 km² in terms of surface area (Ohmura 1987a). The topography reaches an altitude of 3230 m in the center. From there the terrain drops at first slowly and then with increasing steepness toward the sea level. It is steeper in the east than in the west. On the east coast some deep valleys interrupt the otherwise smooth terrain.

All runs were made for an area of 1640 × 2800 km². This includes the entire ice sheet and some parts of the ocean surrounding Greenland. To depict the development of the down slope flow, the wind components and temperatures in several locations have been closely monitored. Four model grid points have been selected for a comparison with measurements. All four represent a different situation regarding the topography. In all areas one automatic weather station is available (Fig. 4.2).

In winter, the boundary layer over the ice sheet is very stable because of the lack of solar radiation. No diurnal cycle is present. The results from the model run are compared with measurements carried out in the month of January. To exclude short-time synoptic changes the monthly-mean daily cycles from the weather stations are compared with the results of a 24 hours period in the simulation.

4.2 Initial conditions

The model begins each run with simplified initial conditions. No attempt has been made to initialize it with real data. The geostrophic wind is zero on all levels. The temperature profile is specified as a simple power-law function of pressure. The major benefit of such a condition is that the hydrostatic equation may be exactly integrated to give an initial analytic geopotential field.

For the topography a data set from the Alfred Wegener Institute (Létrèguilly et al., 1990) has been used (see Fig. 4.1). The grid space is twenty kilometers in both directions, giving 83 grid points in x-direction (positive values are towards the east) and 141 grid points in y-direction (positive values toward the north). To get more stable results the topography has been smoothed and all islands removed. The smoothed terrain is shown in Fig. 4.2.
Figure 4.1: The original terrain computed with the data from the Alfred Wegener Institute. The grid space is 20 km, the total grid points are $83 \times 141$.

Figure 4.2: The smoothed terrain used in the 3-d model. The dots represent the locations of the automatic weather stations (from bottom to top: ETH/CU Camp, GISP2, Tunu, and Humboldt).

The model contains 15 vertical levels with $\sigma$-values of 0.998, 0.99, 0.98, 0.97, 0.96, 0.94, 0.92, 0.90, 0.85, 0.775, 0.70, 0.60, 0.50, 0.30, and 0.10. It is integrated over 72 hours with a time step of 60 seconds. The longwave radiative flux is calculated every 30 minutes and is kept constant during this time. The relative humidity is set at 70% throughout the model. The pressure at the top of the model is 25 kPa.

In the simulation it is assumed that the entire ice sheet is covered with snow. For the Greenland ice sheet the snow density, heat capacity of the snow, thermal condition in the snow, and roughness length of the snow surface are not very well known. Very few measurements at single locations have been carried out during previous field studies, none of them during winter. With this knowledge best guess values had to be chosen. The values
for a run with simplified initial condition are 200 kg m$^{-3}$ for the density, $0.38 \cdot 10^6$ J m$^{-3}$K$^{-1}$ for the heat capacity, and 0.1 W m$^{-1}$s$^{-1}$K$^{-1}$ for the conductivity. For the roughness length a value of $10^{-4}$ m for wind and $10^{-5}$ m for temperature has been taken. These values are assumed to be constant over time and space. The effect of blowing snow on the roughness length has been neglected.

The ocean around Greenland is partially frozen during the whole year. Sea ice was assumed to be covered with snow of the same characteristics as on the ice sheet. In the model a simplified sea ice extension from January 1996 along the coast has been taken; these data were collected by the Institute of Meteorology from the Freie Universität at Berlin. The polar night simulation is not sensitive to the albedo. A constant value of 80 % is set for the snow surface and 5 % for the open ocean. The sea surface temperature for the open ocean is assumed to be -1.9°C.

### 4.3 Wind

The wind vector field at the lowest level is shown in Fig. 4.3. This level corresponds to a height of about 12 m above ground. The wind direction is downslope with a deviation to the right due to the Coriolis force. The wind speed in the interior of the ice sheet is less than 5 m s$^{-1}$. It increases towards the coast, where the slope gets steeper, to more than 10 m s$^{-1}$. The highest wind speed is 10.1 m s$^{-1}$, which has been calculated for the valleys on the east slope. These high values for wind speed are due to a channeling effect, which strengthens the katabatic force (Parish 1984). Over the frozen ocean the wind keeps its direction and weakens slowly. It dies about 300 km from the coast. In the south-western area the wind pattern is different because the ocean is not covered with ice. This leads to a sudden decrease of wind speed over the open water. There, the wind pattern corresponds to the results which were achieved in simulations for similar terrains in Antarctica (Gallice and Schayes 1994).

In the upper left part of Fig. 4.4, the results of wind speed measurements for ETH/CU Camp are being compared with those from the simulation. From the weather station wind data recorded during four years (1992/93/95/96) are available for the month of January. There, the values for the monthly mean wind speed are between 9 m s$^{-1}$ and 11 m s$^{-1}$. Because of the lack of any solar radiation there is no periodic change of any condition. The mean over these four years is almost constant at 10 m s$^{-1}$. The model value is constant over 24 hours at 9.5 m s$^{-1}$. The difference between the individual years is small (less than 0.5 m s$^{-1}$) and the mean over the four years corresponds very well with the results of the model. The upper left picture of Fig. 4.5 shows the wind direction. The results are from measurements for 1994 and 1995. The wind blows from south-east with a direction between 120 and 130 degrees. Both years have similar values and the difference between the measured and the modelled mean is less than 10 degrees.

In the north-western part of Greenland the topography is more complex than in the western part. This area is characterized by two hills which extend into the ocean. The valley between them affects the wind field significantly (Fig. 4.6). The automatic weather station Humboldt lies east of these two hills at the transition zone between the flat interior
Figure 4.3: Wind vectors for the first sigma level after 72 h of winter simulation. For a better visibility only 25 % of the vectors are shown. The grey area represents the part of the ocean which is not covered with sea ice. The four dots represent the position of the automatic weather stations (ETH/CU Camp, Humboldt, Tunu, GISP2). Ice sheet elevation contours are also plotted. The contour interval is 200 m.

and the steep slope. The flat plateau of this area is between 2000 m and 2200 m, and there the wind speed has a value of 4.7 m s\(^{-1}\). Near the coast on the southern side of the valley it has a maximum value of 9.0 m s\(^{-1}\). Within the valley there is a channeling effect of the air masses, which leads to a local wind speed maximum on the southern slope of this valley. Over the frozen ocean the wind weakens only slowly except for a part northward
Figure 4.4: Comparison of the mean daily wind speed measurements with modelled wind speed for January. The results are from the four stations on the Greenland ice sheet and represent a period of 24 hours. ETH/CU Camp top left (results from 92 to 95), Humboldt top right (results from 96 and 97), Tunu bottom left (results from 97) and Gisp2 bottom right (results from 91, 92, 94, and 95). (— model result, - - - individual years, - - - mean)

of the northern hill. This wind pattern influences the potential temperature field, as will be shown later in Fig. 4.16. The strongest potential temperature gradient is calculated for the areas with highest wind speed.

In Fig. 4.7 the wind speed is shown in a cross section from north to south across the two hills. In Fig. 4.10 on the left, four wind profiles from this cross section are shown. There is a different pattern on each slope. On the south slope of the southern hill the maximum wind speed is about 10 m s\(^{-1}\). With increasing distance to the surface the wind speed decreases slowly, reaching a value of almost zero at 1800 m above ground. This part is influenced by air masses flowing down the western slope. On the northern slope of this hill, the wind speed near the surface is higher, namely 12 m s\(^{-1}\) at a height of 100 m. It
Figure 4.5: Comparison of the mean daily wind direction measurements with modelled wind direction for January. The results are from the four stations on the Greenland ice sheet and represent a period of 24 hours. ETH/CU Camp top left (results from 94 and 95), Humboldt top right (results from 96 and 97), Tunu bottom left (results from 97), and Gisp2 bottom left (results from 91, 92, 94, and 95). (-- -- model result, - - - individual years, -- -- mean, 360 deg represents a wind from north)

then decreases to 2 m s\(^{-1}\) at 700 m above ground. The two wind profiles of the northern hill are both from a region with a strongly cooled layer near the surface, as will be shown later in Fig. 4.14. However, the wind speed profiles are different from each other. Above the northern slope there is a layer of 500 m with a strong wind blowing downwards. This layer disappears over the frozen ocean.

The model result for the wind speed at the grid point representing Humboldt station is shown in the upper right of Fig. 4.4. The value decreases slightly from 6.74 m s\(^{-1}\) at the beginning of the 24 hour period to 6.58 m s\(^{-1}\). The monthly means of the daily cycle of wind speed measured at Humboldt in 1996 and 1997 differ only slightly and are almost
Figure 4.6: Wind vector field representing the January model run for the north-western slope of Greenland. The dot represents the position of Humboldt station, the triangles indicate the position of the wind profiles in Fig. 4.10, and the lines stand for the topographic contours with a difference between two lines of 200 m.

constant at 8 m s$^{-1}$ with only a small variation. The difference between model results and measurements done at Humboldt is similar to the one between model results and measurements done on the ETH/CU Camp. The absolute value is different, which is about 2 to 3 m s$^{-1}$ lower. For the wind direction the difference between the two years is about 30 degrees (see upper right of Fig. 4.5), otherwise the model and mean value of the two years correspond very well. The model calculates a direction of 196 degrees.
The region around the Tunu on the north-eastern slope has a very smooth topography. This results in a homogeneous wind field with constant wind speed and direction. A comparison of the measured wind speed with modelled results shows a difference of about 2 m s$^{-1}$ (see lower left part of Fig. 4.4). In the modelled result the wind speed is about 7 m s$^{-1}$ while the measured wind speed is 9 m s$^{-1}$. The wind direction at this point has a difference of less than 20 degrees. The model calculates a wind direction of 275 degrees, the weather station measures approximately 260 degrees.

The situation at the top of the ice sheet is different from the one on the slopes. The wind vector field around the summit shows that the katabatic wind starts almost at the top of the ice sheet. Not far from the summit the katabatic wind is fully developed. For the simulation no geostrophic wind is assumed. Thus, the wind speed at the top is weak, being less than 2 m s$^{-1}$ after 72 hours of simulation (see lower right part of Fig. 4.4). The measured wind speed is influenced by synoptic pressure gradients. Due to this influence the observed wind is stronger than the modelled wind. The mean value is lowest in 1994, with about 4.5 m s$^{-1}$, and highest in 1992, with about 8.5 m s$^{-1}$. The overall mean for the four years is 6 m s$^{-1}$. The measured mean values show that the wind blows from a quadrant between east and south. They further display that in GISP the year to year variation is larger than in the other stations. The model performs a constant shift of the wind direction caused by the higher wind speed on the eastern side of Greenland. This leads to air masses which flow across the top of the ice sheet directed from west to east. This feature builds up slowly and is not yet in a stable state after 72 hours of simulation.

A cross section at the latitude of the summit is shown in Fig. 4.8. It presents the profile of the x-direction of the wind speed, which corresponds to the downslope component of the
Figure 4.8: Cross section of the wind speed in x-direction, which corresponds to the downslope component of the wind speed at the latitude of the summit. Positive values indicate a wind direction to the east and negative values a wind direction to the west. The contour interval is 1 m s$^{-1}$.

Wind after a 72 hour simulation. A maximum wind speed is calculated near the surface within the katabatic layer. At the same altitude the wind is stronger over the steeper eastern slope than over the western slope. Above the surface boundary layer there is a section with no horizontal wind speed in the east-west direction. This section is not on the longitude of the summit but is shifted to the east of it.

Fig. 4.9 shows the cross section for the wind speed in y-direction. At this latitude the y-direction corresponds to the cross slope direction. In the free atmosphere the wind is cyclonic, with a maximum wind speed at an altitude of about 5000 m.a.s.l.; just above the surface it is anticyclonic. This wind pattern causes the airflow to be downward above the summit and upward above the ocean. This type of large scale wind circulation is characteristic for katabatic wind regimes.

Four wind profiles along this cross section are presented in the right of Fig. 4.10. The profile above the steep western slope shows the fully developed katabatic wind with a speed of about 11 m s$^{-1}$ 50 m above ground. However, just above this layer with strong wind there is a steep gradient and the wind decreases to less than 2 m s$^{-1}$ at 600 m above the surface (1800 m.a.s.l). Over the ocean the wind speed near the surface is only 4 m s$^{-1}$ but reaches 2 m s$^{-1}$ at the same height. The two wind profiles at the top of the ice sheet show a thin layer with stronger wind speed. In the upper layers, the profile west of the summit has a stronger wind than the one from the summit.
Figure 4.9: Cross section of the wind speed in y-direction which corresponds to the cross slope wind at the latitude of the summit. Positive values indicate a wind direction to the north and negative values a wind direction to the south. The contour interval is 1 m s\(^{-1}\).

Figure 4.10: Simulated wind profiles at four points along the cross section in north-western Greenland (left part) and central Greenland (right part).
4.4 Temperature

The mean surface temperature for the entire model area is shown in Fig. 4.11. Compared with the temperature distribution map of Ohmura (1987b), the model temperature is about 5°C too warm in the north and about 10°C too cold in the south and in the center of the ice sheet. A comparison with the measurements from 1989 to 1995, however, shows that the simulated temperature is about 5°C too warm at the summit.

Figure 4.11: Mean daily modelled surface temperature for January on the Greenland ice sheet. The dots represent the position of the four automatic weather stations (ETH/CU Camp, Humboldt, Tuna, and GISP2). The ice sheet elevation contours are also plotted. The contour interval is 200 m.
One reason for this difference between the map and model results is that the simulation neglects the energy carried by warmer air from the Atlantic to the south of Greenland. For central Greenland a comparison of the temperature with measurements from the years 1991 to 1995 shows a smaller difference, as will be discussed later.

On the upper left side of Fig. 4.12 the monthly mean daily temperature cycle at the ETH/CU Camp is shown. The difference between the individual years from 1992 to 1995 is almost 10°C, which is rather large. However, the mean over the four years compared with the model output from the first \( \sigma \)-level (about 12 m above the surface) differs less than 3°C. The model calculates a temperature of about -23°C while the measured mean over the four years is -26°C. The model temperature is lower than the warmest year (1992) and is thus within the range of the observed results.

Figure 4.12: Comparison of the mean daily temperature measurements with the modelled temperature from the first \( \sigma \)-level for January. The results are shown for a period of 24 hours at four different stations on the Greenland ice sheet. ETH/CU Camp top left (92 to 95), Humboldt top right (96 and 97), Tunu bottom left (97), and Gisp2 bottom left (91, 92, 94, and 95) (--- model result, - - - individual years, -- -- mean).
For Humboldt the temperature difference between model and measurements is less than 4°C (see upper right side of Fig. 4.12). The measured result is between -35°C and -34°C. The model calculates a temperature of -31°C at the beginning of the period and of -32.5°C at the end. This cooling process can be seen in Fig. 4.17, where the energy terms are shown. The outgoing longwave radiation is only partly compensated by the incoming longwave radiation, and by the sensible heat flux. The remaining energy deficit is responsible for the cooling of the surface layer.

At Tuni (see lower left part of Fig. 4.12) the measured temperature is -36°C. The model temperature corresponds well to the measurements since the maximum difference is less than 2°C.

The comparison of the simulated temperature with the one measured at GISP2 (see lower right of Fig. 4.12) shows that the model calculates a temperature which is slightly warmer. Because of the lower wind speed the turbulent mixing of the surface layer is weaker and there is less energy transported to the cold surface. The model temperature is between -38°C and -40°C. The slow cooling is caused by the energy balance, which is negative for the longwave radiation at the surface. There is no shortwave radiation and the sensible heat flux is less than 1 Wm$^{-2}$.

![Cross section for the potential temperature at 72 degrees north for the January model run. The difference between the isolines is 2°C. The four vertical lines represent the profiles shown on the right side in Fig. 4.14.](image-url)

When starting the model integration period, potential temperature surfaces are aligned in a horizontal manner since geostrophic winds in the free atmosphere are assumed to be zero. The isentropic surfaces quickly move upwards and, as the model integration proceeds, they become oriented in a direction parallel to the surface. By the end of the integration the
isentropic surfaces are nearly parallel to the ice profile. They are packed more tightly above the continental interior and above the frozen ocean than on the steep part of the slope. A cross section at the latitude of the ETH/CU Camp is shown in Fig. 4.13. Notice that nearly adiabatic temperature profiles are found along the ice surface. At a higher altitude in the free atmosphere the potential temperature decreases from east to west.

Four different potential temperature profiles are presented in the right part of Fig. 4.14. They represent the potential temperature over the ocean (80 km from the coast), the slope close to the coast, the steep slope, and the interior of the ice sheet. The strength of the simulated inversion is 12 K over the ocean and in the interior, 8 K on the steep slope, and 2 K near the coast. These values are smaller than the values observed in Antarctica. However, they show the same pattern as the ones for Antarctica, with a weak inversion near the coast and a stronger inversion in the interior (Schwerdtfeger 1984). The wind speed profiles for the same places are shown in the right of Fig. 4.10.

Fig. 4.15 shows the potential temperature field. A cross section of this field along the line marked by triangles is further shown in Fig. 4.16. As with the preceding one, this figure reveals the relative warm air above the northern slopes. This pattern is visible even at 2500 m a.s.l. In the area located near the northern coast, for a deep layer the model simulates the merging of the cold surface layer at a much lower temperature than the rest of the air over the ocean. An analysis of the downslope wind shows that the wind speed decreases simultaneously to the thickening of this cold layer. This indicates the occurrence of an hydraulic jump. The same phenomenon was also observed on the Antarctic coast (Ball 1957; Parish 1981).

The potential temperature is shown in the four profiles on the right side of Fig. 4.14. Two
Figure 4.15: Potential temperature field at the lowest σ-level resulting from the January model run for the north-western slope of Greenland. The dot represents the position of Humboldt station. The triangles represent the position of the profiles displayed in Fig. 4.16. The ice sheet elevation contours are also plotted. The contour interval is 200 m.

Profiles describe the southern hill one for each slope: one profile describes the northern slope of the northern hill; and one the ocean. The profile which represents the southeast point has a constant potential temperature gradient of 0.76 K per 100 m. The profile on the opposite side of the hill has a stronger gradient near the surface. Above this surface layer the gradient is almost zero for 400 m. Then, it takes the temperature of the free
atmosphere. This different state of the atmosphere is also evident in the wind speed profiles (see Fig. 4.10). In the lowest layer the wind speed increases rapidly, while in the southern profile a large layer follows where the wind speed values decrease slowly. In the northern profile the layer is smaller and the gradient steeper. The two profiles representing the northern slope of the second hill and the ocean near the coast both have a strong cooled layer near the surface. However, the wind speed profiles representing these layers are different from each other. Above the slope a strong wind blows downwards, while over the ocean it almost disappears.

Figure 4.16: Cross section for the potential temperature in north-western Greenland after 72 hours of simulation along the line indicating the cross section in Fig. 4.15. The difference between the isolines is 2 K. The vertical lines indicate the position of the profiles shown in Fig. 4.14.
4.5 Energy terms

To observe the development of the katabatic wind it is necessary to isolate the forcing mechanism. In a first step, the net diabatic cooling rate of the katabatic layer is needed. This rate is obtained from the thermodynamic equation and from terms in the equations of motions. These terms are monitored throughout the 72 hours of simulation for five different points. Four of these points represent the automatic weather stations (ETH/CU Camp, Humboldt, Tunu, and GISP2) and a fifth represents the coast (see Fig. 4.3). In the end, the radiative flux divergence at the surface is the driving mechanism for the katabatic flow. The cooling of the air over the slope sets up a pressure gradient force which accelerates the air in downslope direction near the surface. A steady thermodynamic state at the surface is reached when the longwave net radiation is balanced by the turbulent heat flux and by the conduction of energy through the ice.

The time development of the terms in the surface energy budget for five points (ETH/CU Camp, Humboldt, Tunu, GISP2, and at the coast) are illustrated in Fig. 4.17. The net radiation divergence is initially the dominant forcing term; its function is to cool the surface. The turbulent heat flux and the katabatic wind are highly interactive. As the katabatic winds develop, the mixing processes are intensified and the turbulent heat flux toward the surface increases. The turbulent heat flux is strong in the stations with high wind speed, namely ETH/CU Camp, Humboldt, and Tunu. At the coastal station and on the top of the ice sheet the turbulent heat flux is much weaker and the net radiation is balanced by the subsurface heat flux. After 12 hours of simulation the energy terms are balanced and only small changes occur in single terms. These results correspond to the model results for Antarctica, which were obtained by Parish and Waight (1987).
Figure 4.17: From top to bottom: net radiation, sensible heat flux, subsurface heat flux, and energy balance for the surface at five different grid points. The simulation lasted 72 hours and represents the month of January. The locations are: ETH/CU Camp (——), Humboldt station (---), GISP2 station (-- -- --), Tunu (-----), and a coast grid point (---). The place of these locations is shown in Fig. 3.1.
5 Summer simulation

5.1 Introduction

The main focus of this work is the simulation of the summer conditions over the Greenland ice sheet presented in this section. The major difference to the winter conditions is the presence of solar radiation. Also melting or freezing processes become important.

The declination of the sun selected for this simulation corresponds to that in mid July. Comparisons are mainly made between these results and the measurements carried out at the ETH/CU Camp in July 1990 and 1991. The ETH/CU Camp is the only place on the slopes of the Greenland ice sheet where detailed energy measurements have been performed during an entire month. Wind speed, wind direction, and temperature have also been compared with results collected in Humboldt, Tunu, and GISP2. For the most part, observed mean daily cycles for the entire month of July have been compared with the model results. For some parameters the measurements from a sample of days with no clouds have also been taken. Nine cloudless days have been chosen for the ETH/CU Camp. These are the 6th, 7th, 13th, 14th, 17th, 21st, 22nd, 24th, and 25th of July 1991. The observation of the clouds was made during the field season in 1991. For Humboldt, in 1996 there are ten days without clouds, namely the 3th, 6th, 7th, 8th, 12th, 14th, 15th, 17th, 29th, and 30th of July. These cloud observations have been made at Thule, about 250 km south-west of Humboldt, and have been presented in the European Meteorological Bulletin (Deutscher Wetterdienst 1996).

The model run presented in this chapter simulates 66 hours with a 40 km-grid spacing. Longer runs have been made, but the results did not change significantly. The measurements were compared with the result of the final 24 hours of simulation. All results are presented in Universal Time Coordinated (UTC) and True Solar Time (TST). The true solar time difference between the east and the west coast is about three hours and twenty minutes. The TST on the east coast is 1 h 20' behind UTC, on the ETH/CU Camp 3 h 20', and the west coast near Thule 4 h 40'. In this chapter the results are presented either for 6 am UTC and 6 pm UTC or as daily mean values.

5.2 Initial conditions

Many of the initial conditions for the summer simulation are the same as those for the winter simulation described in section 4.2. Thus, only the new initial conditions will be discussed here.

In Greenland part of the coast is not covered with ice. This region, the tundra, is neglected in the model simulation, where the entire land surface is assumed to be covered with ice.

A comparison between the results from Eq. 2.35 in Section 2.2.2 and the measured albedo is shown in Tab. 5.1. The results for the five points compared show a considerable similarity. The albedo distribution for the entire model area is shown in Fig. 5.1. Compared with data measured from satellites, in summer 1990 a similar albedo distribution was found.
The albedo over the open ocean is considered constant at 5% (List 1984).


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The measured mean value for the snow density at the ETH/CU Camp in July 1990 and 1991 is 490 kgm$^{-3}$. This results in a heat capacity of $10^6$ Jm$^{-3}K^{-1}$ and leads to a thermal conductivity of 0.61 Wm$^{-1}$s$^{-1}$K$^{-1}$ (List 1984). These values are used in Eq. 2.43. They are dependent on the height $z$ above sea level. The lack of snow density data for the entire ice sheet, a linear dependence of $K_\beta$ in Eq. 2.43 with the height above sea level is assumed. The function $K_\beta = 130'000 - 30 \times z$ is introduced.

The measurements at the ETH/CU Camp give values between $10^{-3}$ m and $10^{-4}$ m for the roughness length for momentum ($z_{0_M}$) and values between $10^{-4}$ m and $10^{-5}$ m for heat ($z_{0_H}$) (Forrer 1999). For the simulation, a value of $10^{-3}$ m is taken for momentum and one of $10^{-4}$ m for heat.

The ocean is partially frozen during the whole year. For the model run the data on sea ice extension collected in July 1995 by the Institute of Meteorology of the Freie Universität Berlin have been taken.

For the sea surface temperature the mean values for July from the European Center for Medium-Range Weather Forecasts (ECMWF) have been used (see Fig. 5.2). This temperature is held constant during the 66 hours of simulation.
Figure 5.1: Surface albedo for the July simulation (solid lines). Ice sheet elevation contours are also plotted (dashed lines). The contour interval is 200 m.

Figure 5.2: Sea surface temperature for the July simulation (solid lines). Ice sheet elevation contours are also plotted (dashed lines). The contour interval is 200 m.
5.3 Wind

To test the simulation of the katabatic wind in summer the computed surface wind has been compared with the measured wind. The wind field over the entire ice shield is shown in Fig. 5.3 for night conditions (6 am UTC) and in Fig. 5.4 for day conditions (6 pm UTC). The figures show the results from the first $\sigma$-level, which corresponds to a height of about twelve meters above the surface.

Figure 5.3: The wind field from the first $\sigma$-level at night (6 am UTC). The simulation represents the month of July.

At night the wind reaches a maximum speed of 14.6 m s$^{-1}$ on the steep eastern slope of
the ice sheet. The confluence zones, where the cold air from the interior of the ice sheet converges through the valleys to the coast, are a major factor in shaping the character of the katabatic wind above the slope near the coast. The most important confluence zones are on the east coast and north of Thule in the north-western part of Greenland. In most parts of the ice sheet the wind is directed downslope with a deflection to the right due to the balance between pressure gradients, coriolis force and friction. Usually, the katabatic wind reaches the coast but stops blowing immediately over the ocean. In high altitudes the wind speed is low but grows towards the coast.

Figure 5.4: The wind field from the first $\sigma$-level in the afternoon (6 pm UTC). The simulation represents the month of July.
Fig. 5.4 shows the wind field in the afternoon (maximum temperature) from the first \( \sigma \)-level. The pattern does not change significantly from the night. The wind is still directed downwards but with a lower speed (maximum 8.1 m s\(^{-1} \)). Over the ocean there is no change of wind speed or direction except for the east coast. Here the wind from the ice sheet ceases closer to the shore at 6 pm than at 6 am.

The monthly mean daily cycle for wind speed and direction at the ETH/CU Camp are shown in Fig. 5.5. The wind in the model reaches a maximum speed of 10.1 m s\(^{-1} \) at 6 am UTC. This is 1-2 hours earlier than the measured wind speed maximum. The wind speed minimum is 6.0 m s\(^{-1} \) and is reached at 8 pm. This amplitude of 2.1 m s\(^{-1} \) is larger than the one for the measured wind speed, which is 0.7 m s\(^{-1} \). If only the cloudless days are compared, the measured amplitude grows to 1.4 m s\(^{-1} \). This indicates that the daily wind speed cycle is influenced by the intensity of the solar radiation cycle.

The wind direction in the model corresponds well with the measured values. The difference is less than 15 degrees. If only the cloudless days are compared, the maximum difference is 30 degrees. A comparison of the wind direction in the model with the measured wind direction is problematic because the model grid space is 40 km. This means that all topographic patterns which are smaller than 40 km are not represented in the results. In reality though, they can influence the observed direction.

A comparison between simulation and measurements at Humboldt station is shown in Fig. 5.6. The wind speed in the model differs in a maximum of 2.2 m s\(^{-1} \) from the monthly mean over the three years with observations. In addition, there is a difference of 1 m s\(^{-1} \) to results from the sunny days. The wind speed in the model is correlated with the temperature. In the colder morning the wind speed is higher than in the warmer afternoon. This phenomenon is not evident in the measurements. The wind direction of
the model and measured monthly means correspond well with a difference of less than 15 degrees. On the nine cloudless days the difference is similar.

Figure 5.6: Wind speed and wind direction for the month of July at Humboldt: model output (solid lines); monthly mean daily cycle for July 1995-97 (dotted lines); mean over all years (dashed lines); days with no clouds (dashed-dotted lines). 360 deg represents a wind from north.

At Tunu (see Fig. 5.7) the simulated wind speed at noon corresponds well with the observation. At night it differs about 2 m s\(^{-1}\). The maximum difference for the wind direction is less than 30 degrees. The amplitude of the simulated wind speed is about 0.8 m s\(^{-1}\). This is similar to the amplitude at Humboldt but smaller than the one at the ETH/CU Camp.

Figure 5.7: Wind speed and wind direction for the month of July at Tunu: model output (solid lines); monthly mean daily cycle for July 1996 and 1997 (dotted lines); mean over both years (dashed lines). 360 deg represents a wind from north.
The wind profiles in the lowest 1000 m over the ETH/CU Camp are shown in Fig. 5.8. The profiles are presented for noon and midnight. These are the periods when the radio soundings at the ETH/CU Camp were made. At 12 UTC the wind speed near the surface corresponds well. However, in the model it decreases immediately above the first $\sigma$-level. The measured wind profile increases in the first 80 m from 6.4 m s$^{-1}$ to 9.3 m s$^{-1}$ and decreases then to 7 m s$^{-1}$. The higher wind speed above the surface layer represents the geostrophic wind as a result from the synoptic pressure field. The synoptic pressure gradient has been zero at start of the model simulation. During simulation only a weak pressure gradient has been built up above the planetary boundary layer, which is lower than the observed pressure gradient (Agustoni 1993).

![Figure 5.8: Profiles of wind speed at the ETH/CU Camp for 12 UTC and 24 UTC. The profiles from the model calculation (thick solid lines) are split into $x$-components (thick dotted lines) and $y$-components (thick dashed lines). The mean radio sonde profiles from July 1991 (thin solid lines) are split into $x$-components (thin dotted lines) and $y$-components (thin dashed lines).](image)

At 24 UTC the wind speed maximum in the model moves from the lowest level to a height of 50 m. This is lower than in the measurements, where the maximum wind speed is at 150 m above the surface. If the wind is divided into $x$- and $y$-components it becomes evident that the $y$-component, which corresponds to the cross slope wind, is responsible for the change between the two profiles. The layer with strong wind in $x$-direction, which corresponds to the downslope wind, is thicker in the real atmosphere than in the simulation. In the model, the wind turns faster to the $y$-direction than in the measured profile.

A cross section at the latitude of the summit for the mean daily component of the wind in $x$-direction is shown in Fig. 5.9. In Fig. 5.10 the same section is shown for the wind in $y$-direction. At the latitude of these cross sections the wind in $x$-direction corresponds to the upslope direction and the wind in $y$-direction to the cross slope direction.

The wind close to the ice surface blows from the center of the ice sheet towards the coast. In higher latitudes it blows in the opposite direction, near the surface of the ocean towards the coast. The shift to the east of the zero line in the height is a result of the asymmetric shape of the ice sheet.
Figure 5.9: Cross section of the x-component of the wind speed at the latitude of the ETH/CU Camp. Positive values mean wind towards east and negative values wind towards west.

Figure 5.10: Cross section of the y-component of the wind speed at the latitude of the ETH/CU Camp. Positive value means wind towards north and negative values wind towards south.

The cross slope component indicates a cyclonic wind over the sea surface and in the high atmosphere. Over the slope this component shows an anticyclonic circulation. This condition is also evident when the data from all levels are plotted.
5.4 Temperature

The mean daily surface temperature for the entire model area is shown in Fig. 5.11. Compared with the map on the monthly mean surface air temperature from Ohmura (1987b) the model is about two to four degrees colder. The reason for this difference may be that the model calculates the temperature of the surface (snow or ice). In the map, however, the values correspond to the air temperature some meters above the surface. The temperature gradient between surface and screen level may explain some of the differences.

Figure 5.11: Mean daily simulated surface temperature for the month of July for the Greenland ice sheet.
Along the coast where the ocean is not frozen, the temperature in the model remains always at zero degrees. In fact, in the model the surface temperature is not allowed to increase above zero except on the open sea surface.

The daily temperature cycles at the first $\sigma$-level are compared for four different places, the ETH/CU Camp, Humboldt, Tunu, and GISP2 (Figs. 5.12 and 5.13).

![Temperature cycles](image)

Figure 5.12: Mean daily temperature cycle at the ETH/CU Camp (left) and at Humboldt (right) for July: model (solid lines), observation from individual years (dotted lines), observed mean (dashed lines), cloudless days (dashed-dotted lines).

Generally the model represents the measurements well. At the ETH/CU Camp there is a small difference between model results and measurements. However, the shape of the curve is different. After 15 UTC the measured temperature starts to decrease, but, in the model this process does not start before 22 UTC. Also, the minimum value appears in the model about two hours later than in the measurements.

At Humboldt the difference between the observations from the individual years is larger than for the other stations. For this station the model calculates a temperature which is warmer than the measured one. The difference in 1995, the warmest of the three years with observations, is 1.5°C. In 1997, the coldest of the three years with observations, the difference is 5.5°C. A comparison of the nine cloudless days shows an even larger difference of about 7°C. Also at Tunu the simulated temperature is slightly higher than the measured temperature. At night the difference is 3°C and in the afternoon 1°C. At GISP2 the model and the measured temperature coincide well. For all four locations the amplitude of the model temperature is smaller than the measured one.

In Fig. 5.14 the temperature profiles for the lowest 1000 m at the ETH/CU Camp are presented for 12 UTC and 24 UTC. This is the period when the radio soundings were made. The comparison shows a very good agreement between measurements and model. The maximal difference is less than one degree about 200 m above the surface, where the measured temperature is higher than the one in the model. Above 600 m there is again a
maximal difference of less than one degree, with the model calculating higher temperatures than the measured ones.

A cross section of the monthly mean temperature at the latitude of the summit is shown in Fig. 5.15. Above the slopes in the layers near the surface there is an inversion throughout the cross section. By the summit, between the surface and 6000 m a.s.l. the isotherms are on a lower altitude than above the slopes and above the ocean. This indicates descending air masses. Above 6000 m the isotherms above the summit are raising compared to the
surrounding area. This corresponds to the wind pattern shown in Fig. 5.10 with an anticyclonic circulation at lower levels and a cyclonic wind at higher altitudes.
5.5 Energy terms at the surface at the ETH/CU Camp

5.5.1 Shortwave radiation

The diurnal cycle of the shortwave net radiation is presented in Fig. 5.16. The model output is compared with the observed monthly mean daily cycle for July 1990 and 1991. The output is also compared with the mean of the two years and with the mean daily cycle of the nine cloudless days. The observed monthly mean values are 96 Wm$^{-2}$ for July 1990 and 83 Wm$^{-2}$ for July 1991. The mean of these two months is 89 Wm$^{-2}$ (Konzelmann 1994). The model results corresponding to the means of the two years are similar to the measurements with a value of 91 Wm$^{-2}$. For the nine cloudless days there is a difference of 25.3 Wm$^{-2}$ at the time of the maximum shortwave radiation. In the model, the value for the albedo at this grid point is prescribed to be 0.73. The values from the observations are 0.69 for 1990, 0.74 for 1991, and 0.71 for the mean of the nine cloudless days in 1991.

![Figure 5.16: Comparison of the mean daily cycle of the shortwave net radiation between the model output (solid line) and measurements for the month of July at the ETH/CU Camp. The measurements have been divided into the results from July 1990 (upper dotted line) and July 1991 (lower dotted line), the mean of the two years (dashed line) and the measurements of the nine cloudless days (dashed-dotted line).](image)

The modelled mean daily shortwave net radiation for the whole model area is presented in Fig. 5.17. The minimum value of 36 Wm$^{-2}$ is found in the northern part of central Greenland at a height of about 2000 m a.s.l. The value at the south coast reaches 160 Wm$^{-2}$. 
Over the ocean the shortwave net radiation is between 310 and 320 Wm$^{-2}$. These results correspond to the accuracy of albedo measurements. The good agreement between the model results and the measurements done at the ETH/CU Camp has been achieved because of the well-known albedo.

Figure 5.17: The mean daily shortwave net radiation over the Greenland ice sheet (in Wm$^{-2}$) for the month of July. The thick solid line indicates the sea ice edge in July 1995. The values over the open ocean are between 310 and 320 Wm$^{-2}$. 
5.5.2 Longwave radiation

Fig. 5.18 presents the longwave radiation at the surface for the ETH/CU Camp. The modelled longwave net radiation reaches a minimum of \(-77 \text{ Wm}^{-2}\) at noon. The maximum value is \(-59 \text{ Wm}^{-2}\) and is calculated for 6 UTC. Here, there is an amplitude of \(9 \text{ Wm}^{-2}\), which is larger than the \(5 \text{ Wm}^{-2}\) detected in the measurements. The shape of the daily cycle is also different. In the measurements the decrease occurs between 18 and 20 UTC. In the model, however, the shape of the daily cycle starts decreasing earlier. If the values from the cloudless days only are compared, the difference of the mean value disappears; however, the shape is still different. A division of the longwave radiation into the up- and downward components shows on the one hand that the different amplitude in the model results is derived from the outgoing surface radiation; on the other hand, the difference of the mean value stems from the incoming radiation.

Figure 5.18: Comparison of the mean daily cycle of the longwave radiation between the model output (solid lines) and measurements at the ETH/CU Camp. The figure on the left shows the longwave net radiation. On the right the upper figure shows the incoming and the lower figure the outgoing longwave radiation at the surface. The measurements have been divided into the results from 1990 and 1991 (dotted lines), mean of the two years (dashed lines), and the measurements of the nine cloudless days (dashed-dotted lines).

The difference with the incoming longwave radiation can be explained by the fact that the
model does not consider the clouds. Konzelmann (1994) presents for the amount of clouds at the ETH/CU Camp a mean value of 52% for the month of July of 1990 and 1991. As already mentioned, the simulation has been performed with clear sky conditions. Thus, in the model the outgoing longwave radiation is only dependent on the surface temperature.

Figure 5.19: The model output for the mean daily longwave net radiation for July over the Greenland ice sheet (in Wm$^{-2}$).

From 11 to 24 UTC the model surface temperature is 0°C. Therefore, the outgoing long-
wave radiation is constant. The difference between the idealized model surface and the real surface is that in the model the surface temperature is homogeneous. On the ice sheet, though, spots with melting snow alternate with spots where the temperature might be below freezing. This patchwork-like surface might be the reason for the observed curve.

The longwave net radiation at the surface is presented in Fig. 5.19. The values are negative over the entire ice sheet. The lowest value for the longwave net radiation is $-93 \text{ Wm}^{-2}$ and has been calculated for southern Greenland at a height of 2800 m a.s.l. At the summit there is a relative minimum with a value of $-80 \text{ Wm}^{-2}$. From this high altitude to the coast the longwave net radiation increases and reaches values between $-50 \text{ Wm}^{-2}$ and $-40 \text{ Wm}^{-2}$. Over the ocean the values increase with a higher sea surface temperature.
5.5.3 Sensible heat flux

The sensible heat flux at the ETH/CU Camp is presented in Fig. 5.20. It is calculated inside the surface layer from the measured profiles of the 30 m tower by applying the Monin-Obukhov similarity theory (Forrer and Rotach 1997). The curve represents the mean value and the mean of the nine cloudless days for the month of July 1991. The mean values of the model and the ones measured are about the same. In the model, however, the amplitude is about $5 \text{ Wm}^{-2}$ larger. The measured values for the sunny days are similar to the model results in the afternoon but about $5 \text{ Wm}^{-2}$ lower at night.

![Figure 5.20: Comparison of the mean daily cycle of the sensible heat flux between the model output (solid line) and measurements (Wm$^{-2}$) at the ETH/CU Camp (positive values to the surface). The measurements have been divided into the mean for July 1991 (dashed lines) and the mean of the nine cloudless days (dashed-dotted line).](image)

The calculated mean sensible heat flux over the entire ice sheet is shown in Fig. 5.21. The maximum values are found on steep slopes, where the wind speed is high. A maximum value of $52 \text{ Wm}^{-2}$ has been calculated for the steep slopes of south-eastern Greenland. Over the ocean the values are rather low. In the most southern parts, where the sea surface temperature is relatively warm, the value for the sensible heat flux is about $0 \text{ Wm}^{-2}$. This value corresponds well to the results presented by Isemer and Hasse (1987).
Latent heat flux

The latent heat flux at the ground for the ETH/CU Camp is shown in Fig. 5.22. The results from the ETH/CU Camp have been calculated from the humidity profile of the 30 m tower. The difference between the monthly mean and the mean of the nine cloudless days is not significant. In both means the energy flux is almost 0 Wm\(^{-2}\) at 0 UTC and has a minimum of -8 Wm\(^{-2}\) at 12 UTC. The model result is about 10 Wm\(^{-2}\) lower than the measurements, but the shape of the curve corresponds well.

As shown in Fig. 5.23 the values for the mean daily latent heat flux in the interior of the ice sheet and over the ocean are small. On the slopes the value increases to a maximum
Figure 5.22: Comparison of the mean daily cycle of the latent heat flux between the model output (solid line) and measurements. The measurements are split into the results from July 1991 (dashed lines) and the measurements of nine cloudless days (dashed-dotted line) at the ETH/CU Camp (positive values to the surface).

of -22 Wm$^{-2}$. Over the entire ice sheet the values are negative except for a small area on the center of the ice sheet. However, the pattern resembles the one from the sensible heat flux, with the steep slopes producing the highest values.
Figure 5.23: The calculated mean daily latent heat flux for July over the Greenland ice sheet (in Wm$^{-2}$).
6 Discussion and Conclusion

In this work a modified numerical model for katabatic wind has been presented. The model is based on a three-dimensional, hydrostatic, primitive equation, mesoscale model developed by Anthes and Warner (1978), Parish (1984), and Waight (1987). The original model has been extended to simulate the katabatic wind for the entire Greenland ice sheet for any time of the year. To demonstrate the performance of the modified model, results for winter and summer conditions have been compared with measured values from four stations.

In the first part the modified model is described. The governing equations have been adopted from the original model; however, parts of the parameterization of physical processes have been changed. The shortwave radiation at the surface, which is important for the summer simulation, has been parameterized with a formula developed by Konzelmann (1994) and tested with measured values from the ETH expedition. This parameterization leads to good results in cases where the albedo is known. For the latent and sensible heat flux a new parameterization has been chosen and introduced into the model. The parameterization has been tested for values of the ETH/CU Camp by Forrer (1999).

In the second part measurements from four stations situated on the Greenland ice sheet have been discussed. In all these stations the temperature, wind speed, and wind direction have been measured. The observation period extends between one and six years. In addition, at the ETH/CU Camp the energy balance has been measured during two summer seasons. Three of the stations are located on the slopes where the katabatic wind is observed, i.e. in the west (ETH/CU Camp), northwest (Humboldt), and northeast (Tunu) of Greenland. The fourth station (GISP2) is on the summit. In the near future there will be more stations performing continuous measurements. The CIERES has installed a network with fourteen automatic weather stations. Humboldt, Tunu, and ETH/CU Camp are part of this network. Testing of the model could be improved once results from all fourteen stations, particularly from those in the south of Greenland available.

For some important parameters only spare data is available. A significant parameter for a successful summer simulation is the albedo. Measurements from five stations on the ice sheet have been used to parameterize it. For the roughness length, the snow density, and the sensible and latent heat flux measurements from only the ETH/GTT Camp were used to initiate the model and for a comparison with the model results.

A winter simulation has been presented in the third part of this work. The main goal of this simulation was to test whether the model was well adapted to the condition of Greenland. During to the polar night, the situation in Greenland does not differ significantly from the one in Antarctica, for which the model was originally developed. To get better results the topography has been smoothed and all islands along the coast have been removed. The model consists of fifteen levels with a horizontal grid space of 20 km. Tests have shown that a grid space of 40 km leads to very small differences in the results. Therefore, for the summer simulation a grid space with this length has been chosen. The comparison between the measured and the modelled results showed no big difference except for the summit. There the katabatic wind is missing and thus the influence of
The characteristic of the simulated wind corresponds to the observed katabatic wind. The wind direction differs only slightly from the measurements. The modelled wind speed is generally stronger than the measured one. On all three stations on the slope the wind speed maximum during the night is stronger than in the model. The wind speed during the day minimum, however, corresponds well. However, the model is not able to reproduce the diurnal cycle correctly. The synoptic pressure gradient has been neglected for this investigation which might be the reason for some differences between model and measurements.

The simulated temperature is generally warmer than the measured one. A comparison with radio sonde profiles at the ETH/CU Camp shows that the inversion is weaker in the model than in the measured profile. For the calculation of the surface temperature some parameters are not well known. The spatial distribution of the snow density, of the melting and refreezing processes, and of the energy loss due the percolation of water into deeper layers have been parameterized in a rather simple way. To improve this shortcoming, more spatially distributed snow measurements are necessary.

Thanks to the well known albedo, the measured short wave radiation at the ETH/CU Camp is well represented in the model. Without this information there would be a stronger difference between model results and measurements.

The longwave radiation is less well represented in the model because the model calculates the incoming longwave radiation without clouds. The outgoing longwave radiation is dependent on the surface temperature only. An improvement of the surface temperature would ameliorate the outgoing longwave radiation.

The introduction of the latent heat flux and the new parameterization of the sensible heat flux significantly improved the performance of the model for the summer simulation. The parameterization represents well the few available measurements. Still, more measured data is needed to make sure that the chosen parameterization is good enough.

This work has shown that the presented model for the katabatic wind over the Greenland ice sheet represents very well the observed phenomenon. The model is applicable for any time of the year and for any place on the ice sheet. With more measured data and synoptic information the model might be improved to some extent.
A List of Symbols

- $B_v$: Planck function
- $c_p$: specific heat for dry air \([\text{J kg}^{-1} \text{K}^{-1}]\)
- $c_s$: heat capacity for snow \([\text{J m}^{-3} \text{K}^{-1}]\)
- $c_t$: Horizontal diffusion coefficient \([\text{m}^2 \text{s}^{-1}]\)
- $c$: Water vapor pressure \([\text{hPa}]\)
- $c_{\theta_i}$: saturation vapor pressure over ice \([\text{hPa}]\)
- $c_{\theta_w}$: saturation vapor pressure over water \([\text{hPa}]\)
- $E$: phase change information
- $f$: Coriolis parameter
- $F_{u,v,v}$: friction term
- $F^\uparrow$: upward infrared flux \([\text{Wm}^{-2}]\)
- $F^\downarrow$: downward infrared flux \([\text{Wm}^{-2}]\)
- $h$: height of the stable boundary layer \([\text{m}]\)
- $k$: von Karman constant
- $k_f$: friction coefficient
- $K$: Eddy diffusion coefficient
- $K_g$: A constant
- $K_L$: Correction factor \([\text{Wm}^{-2}]\)
- $K_p$: A constant
- $L$: Obukhov length
- $L_v$: latent heat \([\text{J kg}^{-1}]\)
- $m_r$: relative optical airmass
- $p$: atmospheric pressure \([\text{hPa}]\)
- $p_0$: standard surface pressure \([\text{hPa}]\)
- $p_s$: surface pressure \([\text{hPa}]\)
- $p_t$: pressure at the top of the model \([\text{hPa}]\)
- $p^*$: pressure difference between surface and top of the model \([\text{hPa}]\)
- $q$: mixing ratio \([\text{g/kg}]\)
- $Q$: net radiaton \([\text{Wm}^2]\)
- $Q_H$: sensible heat flux \([\text{Wm}^2]\)
- $Q_{LE}$: latent heat flux \([\text{Wm}^2]\)
- $r_E$: radius vector of the Earth \([\text{m}]\)
- $R$: gas constant for dry air \([\text{J kg}^{-1} \text{K}^{-1}]\)
- $R_i$: Gradient Richardson number
- $R_{ib}$: Bulk Richardson number
- $S$: shortwave netto radiation \([\text{Wm}^2]\)
- $S_0$: shortwave radiation at the top of the atmosphere \([\text{Wm}^2]\)
- $S_S$: solar constant \([\text{Wm}]\)
- $t$: time \([\text{s}]\)
- $T$: temperature \([\text{K}]\)
- $T_{\nu}$: monochromatic slab transmission function
- $T_m$: subsurface temperature \([\text{K}]\)
- $T_s$: surface temperature \([\text{K}]\)
### A LIST OF SYMBOLS

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Unit</th>
</tr>
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<tbody>
<tr>
<td>$u$</td>
<td>horizontal component of velocity in $x$-direction</td>
<td>m/s</td>
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<td>$u^*$</td>
<td>friction velocity</td>
<td>m/s</td>
</tr>
<tr>
<td>$U_{i,j,k}$</td>
<td>wind velocity in $x_j$ direction</td>
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</tr>
<tr>
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<td>$V$</td>
<td>wind velocity</td>
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</tr>
<tr>
<td>$z_0M$</td>
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<td>density of water vapor</td>
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<td>kg m$^{-3}$</td>
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<td>geopotential</td>
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A LIST OF SYMBOLS

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<th>Symbol</th>
<th>Description</th>
<th>Unit</th>
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<td>$\Upsilon$</td>
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<td>amount of CO₂</td>
<td>[g/cm²]</td>
</tr>
<tr>
<td>$\Upsilon_t$</td>
<td>amount of absorbing gas above the model</td>
<td>[g/cm²]</td>
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<td>$\Upsilon_w$</td>
<td>amount of water vapor</td>
<td>[g/cm²]</td>
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<tr>
<td>$\Psi_H$</td>
<td>stability function for heat</td>
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</tr>
<tr>
<td>$\Psi_M$</td>
<td>stability function for momentum</td>
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<tr>
<td>$\omega$</td>
<td>angular velocity vector of the earth’s rotation</td>
<td>[radians]</td>
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</table>
B Coordinate transformation

B.1 Basic atmospheric equations

We first summarize the basic atmospheric equations for large-scale flows using Cartesian coordinates x, y, and z directed eastward, northward, and upward (Kasahara 1974).

The equation of horizontal motion may be expressed in the form

\[
\frac{d\vec{v}}{dt} + f(k \times \vec{v}) = -\frac{1}{\rho} \nabla \rho + F,
\]

where

\[
\vec{v} = u_i + v_j, \quad \nabla = i \frac{\partial}{\partial x} + j \frac{\partial}{\partial y}, \quad \frac{d}{dt} = \frac{\partial}{\partial t} + \vec{v} \cdot \nabla + w \frac{\partial}{\partial z},
\]

in which i, j, k denote unit vectors in x—, y—, and z—coordinates, respectively; \(\nabla\) horizontal del operator, \(\vec{v}\) horizontal velocity, \(u\) and \(v\) the x— and y—components of \(\vec{v}\), \(w\) vertical velocity, \(f\) Coriolis parameter \((= 2\Omega \sin \gamma)\), \(\Omega\) angular velocity of the earth’s rotation, \(\gamma\) geographical latitude, \(\rho\) density, \(\rho\) pressure and \(F\) frictional force per unit mass.

For large-scale motions, the hydrostatic equation is

\[
\frac{\partial \rho}{\partial z} = -\rho g,
\]

where \(g\) denotes the earth’s gravity. The mass continuity can be written as

\[
\frac{\partial \rho}{\partial t} + \nabla (\rho \vec{v}) + \frac{\partial (\rho w)}{\partial z} = 0.
\]

B.2 Generalized vertical coordinate system

Let the z—system be the coordinate system with independent variables \(x, y, z, t\), and the \(\sigma\)-system the transformed coordinate system with \(x, y, \sigma, t\), where \(\sigma\) represents the transformed coordinate

\[
\sigma = \sigma(x, y, z, t)
\]

as a function of \(x, y, z,\) and \(t\). We assume that the above equation gives a single-valued monotonic relationship between \(\sigma\) and \(z\), when \(x, y, t\) are held fixed. Thus by inverting B.5 for \(z\), it follows that

\[
z = z(x, y, \sigma, t).
\]

Any scalar function \(A\) in the four-dimensional space may be expressed in either of two ways depending upon whether \(z\) or \(\sigma\) is chosen as a vertical coordinate. Thus, the partial derivation of \(A\) with respect to \(c\), where \(c\) can be \(x, y,\) or \(t\), is generally different in the two systems and the following relationship exists:
where the subscripts $\sigma$ and $z$ indicate a particular vertical coordinate to be held constant for partial differentiation. Using the relationship

$$ \frac{\partial A}{\partial z} = \left( \frac{\partial z}{\partial c} \right) A_{\sigma} \frac{\partial A}{\partial \sigma},$$

we can rewrite B.7 as

$$ \left( \frac{\partial A}{\partial c} \right)_{\sigma} = \left( \frac{\partial A}{\partial c} \right)_{z} + \frac{\partial \sigma}{\partial z} \left( \frac{\partial z}{\partial c} \right)_{\sigma} \frac{\partial A}{\partial \sigma}. \tag{B.9} $$

If we chose $t$ for $c$, it follows that

$$ \left( \frac{\partial A}{\partial t} \right)_{\sigma} = \left( \frac{\partial A}{\partial t} \right)_{z} + \frac{\partial \sigma}{\partial z} \left( \frac{\partial z}{\partial t} \right)_{\sigma} \frac{\partial A}{\partial \sigma}. \tag{B.10} $$

Similarly, choosing $x$ and $y$ for $c$, we obtain

$$ \nabla_{\sigma} A = \nabla_{z} A + \frac{\partial \sigma}{\partial z} \left( \nabla_{\sigma} z \right) \left( \frac{\partial A}{\partial \sigma} \right). \tag{B.11} $$

We now transform the set of prediction equations in the $z$--system into the $\sigma$--system.

**Equation of motion**

The total derivative $d/dt$ in B.2 can be transformed to the $\sigma$-system with the aid of B.10 and B.11. The result is

$$ \frac{d}{dt} = \left( \frac{\partial}{\partial t} \right)_{\sigma} + \vec{v} \cdot \nabla_{\sigma} + \left[ w - \left( \frac{\partial z}{\partial t} \right)_{\sigma} - \vec{\nu} \cdot \nabla_{\sigma} \right] \frac{\partial \sigma}{\partial z} \frac{\partial A}{\partial \sigma}. \tag{B.12} $$

Also, by the definition of total derivative in the $\sigma$--system, we have

$$ \frac{d}{dt} = \left( \frac{\partial}{\partial t} \right)_{\sigma} + \vec{v} \cdot \nabla_{\sigma} + \dot{\sigma} \frac{\partial}{\partial \sigma}, \tag{B.13} $$

where $\dot{\sigma}$ is the transformed vertical velocity

$$ \dot{\sigma} = \frac{d\sigma}{dt} \tag{B.14} $$

which corresponds to $w$ in the $z$--system

$$ w \equiv \frac{dz}{dt}. \tag{B.15} $$

The relationship between $\dot{\sigma}$ can be obtained by comparing B.12 and B.13. Thus,
\[ \dot{\sigma} = \frac{\partial \sigma}{\partial z} \left[ w - \frac{\partial z}{\partial t} \right] - \vec{v} \cdot \nabla_{\sigma} z. \]  
(B.16)

The horizontal equation of motion B.1 can be transformed to the \( \sigma \)-system as
\[ \frac{d\vec{v}}{dt} + f(k \times \vec{v}) = -\frac{1}{\rho} \left( \nabla_{\sigma} p - \frac{\partial p}{\partial \sigma} \frac{\partial \sigma}{\partial z} \nabla_{\sigma} z \right) + \vec{F}. \]  
(B.17)

Transforming the pressure gradient
\[ \frac{1}{\rho} \nabla_{\sigma} p = \sigma \frac{RT}{p} \nabla_{\sigma} p_* \]  
(B.18)

and by making use of the hydrostatic equation B.3 and B.8, we simplify equation B.17 as
\[ \frac{d\vec{v}}{dt} + f(k \times \vec{v}) = -\frac{RT}{p_* + \frac{\epsilon v}{\sigma}} \nabla_{\sigma} p_* - \nabla_{\sigma} \Phi + \vec{F}. \]  
(B.19)

**Continuity equation**

To transform the mass continuity equation B.4 into the \( \sigma \)-system, we have from B.16
\[ w = \left( \frac{\partial z}{\partial t} \right)_\sigma + \vec{v} \cdot \nabla_{\sigma} z + \dot{\sigma} \frac{\partial z}{\partial \sigma}. \]  
(B.20)

Thus,
\[ \frac{\partial w}{\partial z} = \frac{\partial w}{\partial \sigma} \frac{\partial \sigma}{\partial z} = \frac{\partial \sigma}{\partial z} \left[ \frac{d}{dt} \left( \frac{\partial z}{\partial \sigma} \right) + \frac{\partial \vec{v}}{\partial \sigma} \nabla_{\sigma} z + \frac{\partial \dot{\sigma}}{\partial \sigma} \right]. \]  
(B.21)

Also with the aid of B.9, it can been shown that
\[ \nabla_{\sigma} \cdot \vec{v} = \nabla_{\sigma} \cdot \vec{v} - \left( \frac{\partial \sigma}{\partial z} \right) \nabla_{\sigma} z \frac{\partial \vec{v}}{\partial \sigma}. \]  
(B.22)

Substitution of B.21 and B.22 into B.4 yields the mass continuity equation in the \( \sigma \)-system:
\[ \frac{d}{dt} \ln \left( \rho \frac{\partial z}{\partial \sigma} \right) + \nabla_{\sigma} \cdot \vec{v} + \frac{\partial \dot{\sigma}}{\partial \sigma} = 0 \]  
(B.23)

or
\[ \frac{d}{dt} \left( \rho \frac{\partial z}{\partial \sigma} \right) + \left( \rho \frac{\partial z}{\partial \sigma} \right) \left( \nabla_{\sigma} \cdot \vec{v} + \frac{\partial \dot{\sigma}}{\partial \sigma} \right) = 0. \]  
(B.24)

Substitution of B.13 in B.24 yields
\[ \frac{\partial}{\partial t} \left( \rho \frac{\partial z}{\partial \sigma} \right) + \nabla_{\sigma} \left( \rho \vec{v} \frac{\partial z}{\partial \sigma} \right) + \frac{\partial}{\partial \sigma} \left( \rho \dot{\sigma} \frac{\partial z}{\partial \sigma} \right) = 0. \]  
(B.25)
Introducing the hydrostatic equation and with the aid of B.8

\[ \frac{\partial}{\partial t} \left( \frac{\partial p}{\partial \sigma} \right) + \nabla \cdot \left( \frac{\partial \underline{p}}{\partial \sigma} \frac{\partial \vec{v}}{\partial t} \right) + \frac{\partial}{\partial \sigma} \left( \frac{\partial \underline{p}}{\partial \sigma} \frac{\partial \vec{v}}{\partial t} \right) = 0. \] (B.26)

With \( \frac{\partial p}{\partial \sigma} = p_* \)

\[ \frac{\partial p_*}{\partial t} = -\nabla \cdot (p_* \vec{v}) - \frac{\partial p_* \dot{\sigma}}{\partial \sigma}. \] (B.27)

With the integration from the surface to the top of the model and with the boundary condition \( \dot{\sigma} = 0 \) at \( \sigma = 0 \) and at \( \sigma = 1 \) the continuity equation B.4 can be transformed to the \( \sigma \)-system

\[ \frac{\partial p_*}{\partial t} = -\int_0^1 \left( \frac{\partial p_* \vec{u}}{\partial x} + \frac{\partial p_* \vec{v}}{\partial y} \right) d\sigma. \] (B.28)

**Hydrostatic equation**

The hydrostatic equation B.3 in the \( \sigma \)-system can be expressed, using B.8, as

\[ \frac{\partial \rho}{\partial \sigma} = -\frac{1}{g} \frac{\partial p}{\partial \sigma}, \] (B.29)

and with \( d\Phi = g dz \)

\[ \frac{\partial \Phi}{\partial \sigma} = -\frac{\partial p}{\partial \sigma}. \] (B.30)

Also, with the aid of \( p = \sigma p_* + p_T, \frac{\partial p_T}{\partial \sigma} = 0 \), and \( \frac{\partial p_*}{\partial \sigma} = 0 \) we have

\[ \frac{\partial p}{\partial \sigma} = p_* . \] (B.31)

Substitution of B.30 and B.31 in B.29 and with \( p = \rho RT \) leads to

\[ \frac{\partial \Phi}{\partial \sigma} = -\frac{RT}{\sigma p_* + p_T} p_* . \] (B.32)

Rearranging these equation leads to the form of the hydrostatic equation used in the model

\[ \frac{\partial \Phi}{\partial \ln \left( \sigma + \frac{\epsilon \tau}{p_*} \right)} = -RT . \] (B.33)
Equation for the vertical velocity

The integration of (B.27) from the top of the model to the actual height leads to

\[ \int_{\sigma}^{1} \left( \frac{\partial p_*}{\partial t} + \nabla_{\sigma}(p_* \vec{v}) \right) d\sigma - p_* \dot{\sigma} = 0, \]  

and to the form used in the model

\[ \dot{\sigma} = -\frac{1}{p_*} \int_{1}^{\sigma} \left( \frac{\partial p_*}{\partial t} + \nabla_{\sigma}(p_* \vec{v}) \right) d\sigma. \]
References


REFERENCES


REFERENCES


Curriculum Vitae

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