

On the Arctic Boundary Layer

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From Turbulence to Climate

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Frontpage image: As a fresh PhD. student I visited Uppsala University January 2003 to attend a seminar by Larry Mahrt. In advance I decided to contact Professor Sergej Zilitinkevich for a discussion after the seminar. Of course, Sergej had forgotten everything about our appointment, so I had to wait around for several hours. Finally, after less than thirty minutes discussion and a piece of paper filled with cryptic notes (the image) the directions were set for a wonderful collaboration. On the right are the basic assumptions utilized in Paper III. Photo by Malin Mauritsen.

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Abstract

The boundary layer is the part of the atmosphere that is in direct contact with the ground via turbulent motion. At mid-latitudes the boundary layer is usually one or a few kilometers deep, while in the Arctic it is much more shallow, typically a few hundred meters or less. The reason is that here the absolute temperature increases in the lowest kilometer, making the boundary layer semi-permanently stably stratified. The exchange of heat, momentum and tracers between the atmosphere, ocean and ground under stable stratification is discussed from an observational, modeling and climate-change point of view. A compilation of six observational datasets, ordered by the Richardson number (rather than the widely used Monin-Obukhov length) reveals new information about turbulence in the very stably stratified regime. An essentially new turbulence closure model, based on the total turbulent energy concept and these observational datasets, is developed and tested against large-eddy simulations with promising results. The role of mesoscale motion in the exchange between the atmosphere and surface is investigated both for observations and in idealized model simulations. Finally, it is found that the stably stratified boundary layer is more sensitive to external surface forcing than its neutral and convective counterparts. It is speculated that this could be part of the explanation for the observed Arctic amplification of climate change.

List of Papers

This thesis is based on the following papers, which are referred to in the text by their Roman numerals:

- I Mauritsen, T., G. Svensson and B. Grisogono, 2005: Wave flow simulations over Arctic leads. *Boundary-Layer Meteorology*, **117**(2), 259-273.
- II Mauritsen, T. and G. Svensson, 2007: Observations of stably stratified shear-driven atmospheric turbulence at low and high Richardson numbers. *Journal of the Atmospheric Sciences*, **64**(2), 645–655.
- III Mauritsen, T., G. Svensson, S. S. Zilitinkevich, I. Esau, L. Enger and B. Grisogono, 2007: A total turbulent energy closure model for neutral and stably stratified atmospheric boundary layers. *Submitted to Journal of the Atmospheric Sciences*.
- IV Tjernström, M. and T. Mauritsen, 2007: Mesoscale variability in the summer Arctic boundary layer. *Manuscript*.
- V Mauritsen, T. and G. Svensson, 2007: Sensitivity of the dry stable boundary layer to external surface forcing. *Manuscript*.

Reprints were made with kind permission of Springer Science and Business Media and the American Meteorological Society. I have been the main contributor and written most parts of Papers I, II, III and V. The original ideas for Papers I and III belong to Gunilla Svensson and Sergej S. Zilitinkevich, respectively. I had the original idea of analyzing the micro-barograph data from AOE-2001 for Paper IV, but insufficient familiarity with the large observational dataset. Therefore, Michael Tjernström has done the major part of the manuscript.

Publications not included in the thesis:

- Cuxart J., A. A. M. Holtslag, R. Beare, A. Beljaars, A. Cheng, L. Conangla, M. Ek, F. Freedman, R. Hamdi, A. Kerstein, H. Kitagawa, G. Lenderik, D. Lewellen, J. Mailhot, T. Mauritsen, V. Perov, G. Schayes, G-J. Steeneveld, G. Svensson, P. Taylor, S. Wunsch, W. Weng and K-M. Xu, 2006. Single-column intercomparison for a stably stratified atmospheric boundary layer. *Boundary-Layer Meteorology*, **118**, 273–303.
- Mauritsen, T. and E. Källén, 2004. Blocking prediction in an ensemble forecasting system. *Tellus*, **56A**, 218-228.
- Steeneveld, G. J., T. Mauritsen, E. I. F. de Bruijn, J. Vila Guerau de Arellano, G. Svensson, and A. A. M. Holtslag, 2007. Evaluation of the stable boundary layer and diurnal cycle in the limited area models MM5, COAMPS and HIRLAM for three contrasting nights in CASES99. *Submitted to Journal of Applied Meteorology and Climate*.

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1. Introduction

The Earth's northern polar region is named the Arctic. It is the home of a unique wildlife, including the Polar bear, and highly adapted indigeneous peoples. It is also one of the least explored and exploited regions on Earth by mankind, due mainly to the immense practical problems with accessing the region. All of this appears to be changing. The ongoing dramatic climate changes in the Arctic region has recently received great attention in both the scientific community and in the public media. Alarming reports of sea ice, snow and glaciers melting and thawing tundra, rising sea-levels, reductions of the Gulfstream and changes in vegetation and wildlife are flooding the press. Indeed the Arctic region has warmed more than anywhere else on Earth in the past century; a phenomenon denoted polar amplification. The historic warming rate is roughly twice as fast as the global average.

We have long been fascinated by the Arctic. Through time numerous explorers have attempted to unveil its secrets and many of them have paid with their lives. Modern exploration has been conducted since the fifteenth century, though we believe the Greeks sailed into the North Atlantic, visiting for example Iceland during their reign. In historical times the explorations were initiated by whale hunters. These were soon to be followed by explorers with higher goals, such as finding a route to the Pacific Ocean and even later the motivation was the pure glory of being the first to reach the North Pole. In 1893 the Norwegian explorer and oceanographer Fridtjof Nansen made a serious attempt at reaching the pole. He and his crew froze their ship Fram into the pack ice north of Siberia. Nansen was convinced that the ocean currents would bring them across the Arctic Ocean towards Svalbard and Greenland. His predictions were correct; Fram drifted across the ocean and reached Norway three years later. At some point it became obvious that they would not pass the pole. Determined to reach his goal, Nansen and his companion went off on foot, though they were unsuccessful due to the harsh conditions. They barely survived on Franz Josefs land and were ultimately rescued by other explorers. The first to reach the Pole were the Americans Robert Peary and Matthew Henson. They used the knowledge and skill of the Inuit people on dogsleds, furs, and igloos to achieve their goal. They went from Ellesmere Island to the Pole in just little more than a month in the spring of 1909. Today scientists and wealthy tourists frequently visit the Central Arctic Ocean aboard modern icebreakers.

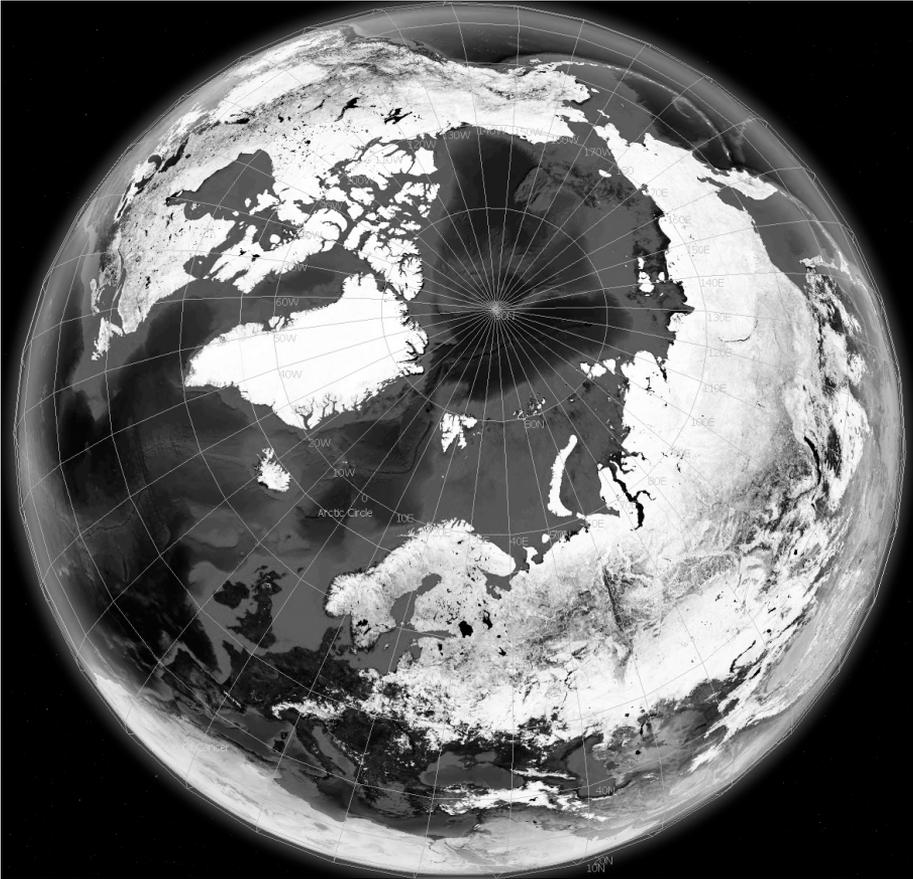


Figure 1.1: A satellite composite for January obtained from NASA World Wind.

Today, we know that the Arctic region consists of a vast ice-covered ocean surrounded by continents. The Arctic ocean is connected to the Atlantic Ocean through the Nordic Seas and the Canadian Arctic Archipelago as well as to the Pacific Ocean through the relatively narrow and shallow Bering Strait (Figure 1.1). Various definitions exist of the extent of the Arctic. One definition is the area north of the Arctic Circle (67°N), which limits the area with polar night and midnight sun. The climatic and ecological definition is the area with a mean July temperature below 10°C , which roughly corresponds to the tree-line. The Arctic as defined by this mean is not constant in time. Another practical definition is everything north of 60°N , which includes some of the sub-arctic. Even with this rather wide definition, the Arctic region only comprises 6.7 percent of the Earth's surface. Consequently, the rapid changes occurring in the Arctic have only little impact on the global average temperature due to the negligible size of the region. On the other hand, the local impacts are tremendous on the unique fauna, wildlife and people.

Conditions in the Arctic are very different from the rest of the globe. The environment offers challenges to its inhabitants, as well as to scientists trying to understand it. There is the sea-ice and tundra on land, clouds appear to be different, the dark polar night during winter and the midnight sun during summer to mention a few. A major theme in the present thesis is the consequences of the atmosphere often being more 'layered' than elsewhere on the globe. Due to the low temperatures at the surface compared with higher up, air is reluctant to mix vertically, which can result in multiple thin layers of clouds and pollutants. We refer to the atmosphere as being stably stratified. The lowest kilometer of the Arctic atmosphere is almost permanently stably stratified year around, more so during winter than summer.

Scientists in the 1980's warned that the Earth's climate was warming, after a period when some researchers claimed that we were heading towards a new ice-age. In order to sort things out the Intergovernmental Panel on Climate Change (IPCC) was jointly established by the World Meteorological Organization (WMO) and the United Nations Environment Programme (UNEP) in 1988. IPCC has provided us with three scientific assessment reports and the fourth is to appear by the time of printing of this thesis. In the third assessment report it is concluded that *'The observed warming in the latter half of the 20th century appears to be inconsistent with natural external (solar and volcanic) forcing of the climate system'* and continues *'Anthropogenic factors do provide an explanation of 20th century temperature change'*. The anthropogenic factors referred to are man-made emissions of greenhouse gases and sulphate aerosols. Note that IPCC does not exclude alternative explanations of the warming. The foundation of modern science is that no theory can be proven right, it can only be disproved. Failing to disprove a theory despite strong efforts, increases our confidence in it. It is important to distinguish the scientific from the inevitably associated political processes.

In addition to the investigation of the past climate changes, IPCC conducted a number of projections of future climate using the best available climate models. The conclusion is that the Earth may warm by 2° to 5°C if the atmospheric carbon dioxide concentration is doubled. Most projections of man-made changes indicate larger concentration increases. These numbers are put into perspective by ice-core based estimates that the last glacial minimum was only locally 7°C colder than present climate in Antarctica (e.g. Watanabe et al. 2003). Projections of future climate indicate that the observed polar amplification will continue. Climate models indicate that the Arctic region will warm at rate between 1.4 and 4 times faster than the global average (Bony et al. 2006). The uncertainty in the Arctic is larger than elsewhere. For example some models project ice-free ocean summers by the end of the century while other show virtually no change in ice extent. In order to both narrow down this uncertainty and to evaluate the consequences of polar amplification on nature and society the Arctic Council and the International Arctic Science Committee launched the Arctic Climate Impact Assessment (ACIA) project,

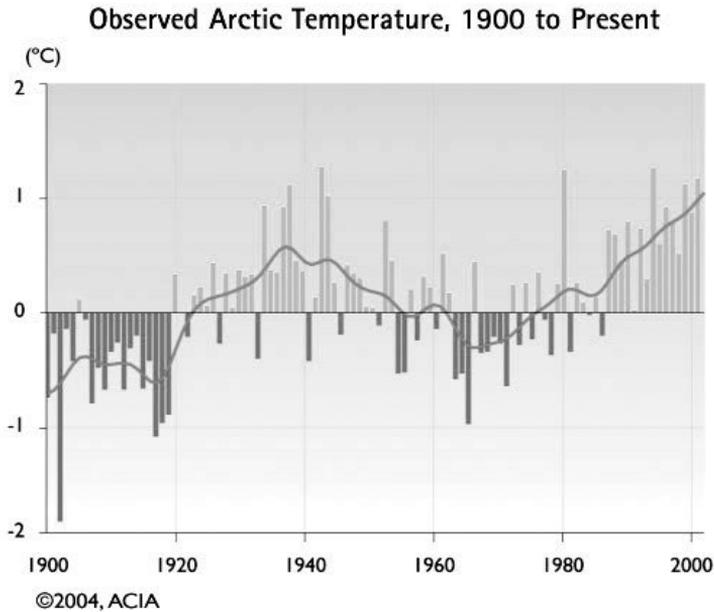


Figure 1.2: Observed Arctic temperature averaged over the area north of 60°N. Figure courtesy granted by Paul Grabhorn, ACIA.

which published a comprehensive scientific report in 2005. Figure 1.2 shows a plot from the report (ACIA, 2005), showing the observed temperature in the Arctic.

In principle the fundamental physical processes that together govern the Arctic climate system are the same over the entire globe and probably even on other planets. Clouds-droplets form on aerosols both in the Arctic and lower latitudes, a turbulent eddy doesn't know where it is on the planet, unless it is very large and feel the Coriolis force, a photon behaves according to quantum mechanics regardless of latitude. So how come we have such difficulty in assessing the response of the Arctic climate system? Maybe we have been pushed outside the envelope of our understanding when faced with Arctic climate? Things we thought were unimportant suddenly turn out to be relevant, and consequently the emergent behaviour of the complex climate system is different from what it is at the familiar mid-latitudes.

1.1 The Arctic Climate System

Climate is the mean and variability of the weather averaged over a relatively long period¹. We usually refer to near-surface variables, such as the temperature, wind and precipitation when describing the climate. The Earth's climate

¹The World Meteorological Organisation has defined the averaging period to be 30 years.

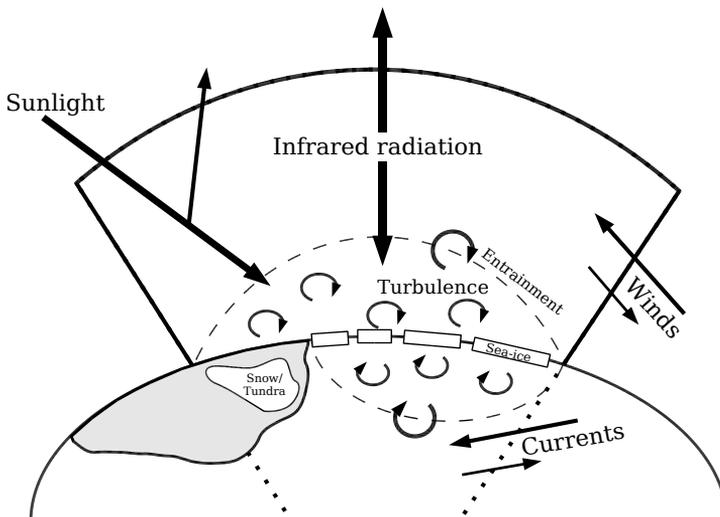


Figure 1.3: Illustration of the Arctic climate system. The two dashed curves depict the Arctic inversions.

is determined by the interaction between a number of components; the atmosphere, ocean, cryosphere, biosphere and lithosphere. We refer to these components as constituting the climate system and, strictly speaking, climate is the statistical state of the climate system. The exchange between the components occurs over widely different timescales. For instance, responses related to the lithosphere occur over thousands to billions of years, which is why we often neglect them in studies of the near-future climate change.

When we consider global climate, the limit of the study is conveniently set at the top of the atmosphere. This limits the exchange of energy in and out of the system to radiation. Sunlight is partially absorbed by the atmosphere, land-surface and ocean, and partially reflected back to space. The climate system emits infrared radiation, depending on its effective temperature. Taken over sufficiently long periods of time, the energy should balance between incoming and outgoing energy at the top of the of the atmosphere. In regional climate studies we need to consider a box with exchange with the rest of the Earth through the sides. Figure 1.3 illustrates how this works for the Arctic. In the Arctic region, there does not exist a simple balance of radiation at the top of the Arctic atmosphere, rather there is a radiative deficit between incoming solar radiation and outgoing infrared radiation. The Arctic is cold, though warmer than it is supposed to be. Atmospheric winds and ocean currents provide exchange through the sides of the Arctic box, and since the Arctic is cold compared with lower latitudes, there is a net transport of heat to the Arctic restoring the balance.

The emitted infrared radiation to space is proportional to the fourth power of the Earth's effective temperature. It is the temperature Earth would have if it had no atmosphere and acted like a black body, absorbing all the incoming radiation received at its surface and reradiating it all back to space. The effective temperature of the Earth as a whole is about -17°C . Had it not been for the atmosphere's greenhouse effect the planet would have this average temperature. In this scenario the temperature would vary wildly, by hundreds of degrees, between day and night, summer and winter. Popularly speaking, the greenhouse effect warms the lower parts of the atmosphere by absorbing some of the infrared radiation and emitting it in all directions, including back toward the surface. Water vapor is the most important greenhouse gas, as it accounts for about one third of the total greenhouse effect. Other important greenhouse gases are carbon dioxide, ozone, methane and nitrous oxide. In addition clouds, which mostly consist of water, are important absorbers of infrared radiation. The main components of the atmosphere, molecular nitrogen, oxygen and argon, are not greenhouse gases.

If we consider the Arctic surface temperature things get more complicated. Here a large number of physical processes are at play within the atmosphere, ocean, sea-ice and ground. The major components are the incoming and reflected solar radiation, the downward infrared radiation emitted from the atmospheric greenhouse gases, infrared radiation emitted by the surface, conducted heat from the ocean, sea-ice, snow, tundra and the ground, and finally, the exchange of heat to and from the atmosphere by atmospheric motion. The latter is a main topic in the present thesis. Exchange at the surface by atmospheric motion is often referred to as turbulent exchange. It consists of two components; the sensible heat flux is caused by mixing of air with different temperatures and is directed from warm to cold, while the latent heat flux consists of the latent heat of evaporation usually directed from the surface to the atmosphere. The latent heat flux is very small in the Arctic (Persson et al. 2002) because the atmosphere can hold very little water at low temperatures.

Exchange between the atmosphere and the surface has to overcome the Arctic temperature inversion, one of the major features of the Arctic atmosphere. It is generated as relatively warm air from the midlatitudes moves into the Arctic region aided by low-pressure systems. Here the intruding warm air slowly rises, or slides up, over the cold and dense air close to the surface. The latter is kept cold by the radiative deficit. In wintertime, the temperature is observed to increase with height on average by roughly 10°C in the lowermost kilometer (e.g. Kahl et al. 1996). In a well-mixed dry atmosphere the temperature would have decreased by the same amount, due to the adiabatic cooling with height. A similar inversion is present in the uppermost ocean, where relatively fresh and cold water resides in a layer of 50-100 meters above the saline, warm, and hence denser water of North Atlantic origin.

The two Arctic inversions effectively shield the Arctic surface from the free atmosphere and deep ocean. The reason is that the air is stably stratified: A

small parcel of air from the free atmosphere is potentially less dense than a small parcel close to the ground. By 'potential density' we mean the density given the same pressure without exchange of heat with the surroundings (an adiabatic process). If we were to try to move the parcel from the free atmosphere into the boundary layer we would feel a positive buoyancy force, which would attempt to move the parcel back to its original position. In other words; it takes energy to mix over an inversion. On the other hand, the inversion is there because there is a lack of, or only very weak, mixing across it. Vertical exchange of heat in the Arctic is in sharp contrast to the tropics, where a slight warming of the surface will result in a warming throughout the troposphere. Deep convective cells convey the heat away from the surface in impressive thunderstorms. In fact, the warming in the upper parts of the troposphere will be larger than at the surface, due to the simultaneous release of latent heat. We shall return to these differences between the Arctic and the rest of Earth in Chapter 5, where the issue of polar amplification is addressed.

1.2 Boundary Layers and Turbulence

The Earth's atmosphere is traditionally divided into a number of layers in the vertical. The lower part, which is denoted the boundary layer, is loosely defined as being influenced by the surface on a timescale of an hour. For example, if the surface temperature were to change it would be felt within the entire boundary layer in roughly one hour. Typically, the boundary layer is said to encompass the lowest kilometer of the atmosphere, a value often cited in textbooks. However, the depth of the boundary layer may vary from just a few meters during calm clear nights up to several kilometers on sunny summer days.

Once in a while, we can observe the boundary layer without a need for sophisticated instrumentation. One such example is the formation of shallow fogs around sunset. Here the evaporating moisture from the ground is trapped within a very thin layer, while at the same time the temperature drops, such that the humidity reaches saturation. These shallow and stably stratified boundary layers, not the fog itself, are the main focus of this thesis. Another prominent example is the fair-weather cumulus clouds that form in the afternoon over land in the upper part of the boundary layer. Here parcels of warm and moist buoyant air rise from the surface as they are less heavy than their environment. As they rise, they cool due to the falling pressure with height, such that at some point a small cloud may form.

Within the boundary layer the flow is dominated by turbulence. Most people know turbulence from commercial flights, where it is referred to as 'holes in the air'. This labelling is a misconception. In fact the atmosphere contains no voids, rather the sensation of falling into a hole is caused by small downward windgusts. The human fear of falling spawns an immediate response,

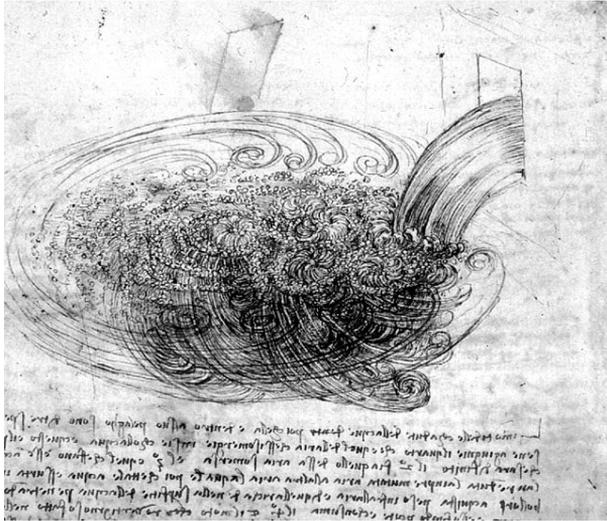


Figure 1.4: Drawing by Leonardo da Vinci of turbulent eddies generated by water flowing into a basin. The drawing was obtained from <http://www.visi.com/reuteler/>.

even though the drop of commercial aircrafts seldom exceeds one meter. Turbulence is also responsible for the sound of blowing wind and the rapid dispersion of odours. However, the most important influences of turbulence on our lives are invisible to the naked eye.

Turbulent flows, governed by the fundamental laws of physics and well described by the laws of mechanics, remain one of the fundamental unsolved problems in physics. Most scientist even agree that it is probably an unsolvable problem, at least within our present understanding.

The modern theories and discoveries of turbulence were made for the most part more than fifty years ago in Russia by prominent scientist such as Landau, Kolmogorov and Monin to mention a few. Earlier scientific contributions from Reynolds, Richardson and Prandtl should be included. Their findings in turn build on the work of Newton, Bernoulli and Euler. Possibly they found their inspiration in the artistic drawing by Leonardo da Vinci, which date around the year 1500 (Figure 1.2). We can only speculate on which insights Leonardo possessed concerning the nature of turbulence. The parodic poem² by Lewis Fry Richardson from around 1920 nicely illustrates the fractal nature of turbulence:

"Big whorls have little whorls
that feed on their velocity;

²The poem is based on the Anglo-Irish priest and satirist Jonathan Swift's poem on infinity: 'So, nat'ralists observe, a flea; Hath smaller fleas that on him prey; And these have smaller yet to bite 'em; And so proceed *ad infinitum*; Thus every poet, in his kind, Is bit by him that comes behind.'

and little whorls have lesser whorls
and so on to viscosity."

The largest turbulent eddies are produced by, for example, the flow around an obstacle. In this case the size of the eddies is determined roughly by the size of the obstacle. In da Vinci's drawing the largest eddy is probably limited by the size of the basin in combination with the size of the water inlet. These big eddies are then broken into gradually smaller and smaller eddies until finally dissipated by molecular viscosity at the millimeter scale. Note that in da Vinci's drawing each eddy has a curly tail with another smaller eddy on it. The difficulty of dealing with turbulence lies in the interaction of widely different scales of motion. Richardson's poetic observation was later to manifest itself in the scientific community by the aid of Kolmogorov's (1941) probabilistic theory on turbulence. The original papers by Kolmogorov are difficult to grasp even for scientists working in the field. Most of us rely on interpretations by others, rather than reading the original texts ourselves (e.g. Frisch, 1995).

As vague as the theories on turbulence may be, equally hazy is the application of these theories to practical problems. Each situation is unique and requires special treatment in the form of empirical assumptions. In fact it is possible to demonstrate that problems involving turbulence cannot be solved without assumptions. This is known as the turbulence closure problem. A large number of different kinds of assumptions exists in the literature, and it would be futile to review them all here. However, a very popular choice is to define a typical turbulence length-scale for the problem at hand. The length-scale defines the size of the largest turbulent eddies. For example, if we consider turbulence close to a fixed barrier, we can quite easily imagine that the turbulent eddies cannot be much larger than the distance to the barrier. This is known as *the law of the wall*, which ought to be used in every model of atmospheric turbulence close to surface. Some distance away from the surface other length-scales dominate the flow.

Applying an incorrect or inaccurate turbulence length-scale or an otherwise erroneous turbulence model to a problem leads to degraded solutions. This is the case for both theoretical work and for the models used in weather prediction and climate research. Under the Arctic inversion turbulent mixing is a source of heat to the surface because it will diffuse warm air down from above. If the model we use for turbulent exchange is too diffusive under such conditions, the surface temperature will be biased high. Imagine that we are in a place where the observed climatic mean temperature is 15°C. Our erroneous climate model tells us that it is 20°C and in the future it will increase to 23°C. We can then compensate for the known bias of 5°C and conclude that in the future the temperature is likely to be 18°C. This is a quite uncritical compensation. However, things get more complicated when we consider a place like Siberia or Alaska where the mean temperature is just below zero. These regions of tundra with permafrost relies on the annual mean temperature being

below zero. Our erroneous model would not have the tundra in first place. Or consider the sea-ice extent in a model that has such a bias. Unfortunately, a bias of 5°C is not uncommon in climate models and for the Arctic region, it is often much larger.

1.3 Major Questions Addressed

This thesis is based on the papers I-V listed in the beginning. Out of necessity, these are written at a scientific level which is difficult to approach for most readers and they further contain many details of little interest to anyone but scientists in the specific fields. I therefore decided to let the following four chapters treat general problems of meteorology, all of which can be found in textbooks in use for under-graduate and graduate level teaching. Obviously, papers I-V do not cover the width and breadth of these fundamental questions, rather, they should be viewed as small contributions to the respective scientific fields.

1. The most important atmospheric motions, that on average contain the most energy, are the large-scale synoptic, or weather, and the small-scale turbulent boundary layer scales. Each scale has a certain amount of energy associated with it. It is fundamental to many present-day observational techniques and to many atmospheric models that the synoptic- and boundary layer scales are separated by a spectral energy 'gap' at the intermediate mesoscales. The problem, and the existence of such a gap is discussed in Chapter 2.
2. Turbulence under stable stratification has long been thought to decay beyond a certain limit, the critical Richardson number (Richardson, 1920). Here the stratification becomes so strong that turbulent eddies will not be able to exist. Chapter 3 brings forth new evidence that this picture needs to be refined. Rather than a non-turbulent state, super-critical turbulence appears weak, though active.
3. Modeling of atmospheric turbulence under stable stratification is a major problem in numerical weather forecasting and climate simulations. It is most often based on work by Obukhov (1946), where the vertical turbulent fluxes of momentum and heat are described by stability functions of the turbulent fluxes themselves. An alternative approach is sketched in Chapter 4, where instead the gradient Richardson number is used.
4. Historic observations and climate model simulations indicate that the Arctic region is more sensitive to external climate forcing than the rest of the Earth, the Arctic amplification phenomenon. A common hypothesis explaining Arctic amplification is related to the decreasing reflectivity of the surface in a warming scenario as snow and ice melts. Chapter 5 questions the importance of this hypothesis

and widens the discussion to include the effect of stable stratification close to ground on climate sensitivity.

2. Atmospheric Scales of Motion

Atmospheric motion occur on all scales, with an upper limit dictated by the size of Earth and a lower given by the molecular viscosity of air. In practice this means scales from millimeters to tens of thousands of kilometers. A comprehensive understanding of the atmosphere involves some kind of treatment of all these scales of motion, including their interaction. We usually divide the atmospheric motions into different categories as discussed in Paper IV: Planetary, synoptic, mesoscale and boundary-layer scales. Planetary scale motions of a magnitude which is a fraction of the Earth's circumference and are closely linked with the temperature and pressure distributions. Therefore, they are usually characterized by their planetary or hemispheric wavenumbers; for instance motions with a positive pressure anomaly in Asia and negative in North America is at wavenumber one. Another example is the Arctic Oscillation which, roughly speaking, is the pressure difference between the high- and the mid-latitudes. The synoptic scales are the typical sizes of low- and high-pressure weather systems (Bjerknes and Solberg, 1922; Holton, 2004). The boundary-layer scales range from small-scale turbulence to eddies over the depth of the boundary layer. The mesoscales, meaning the scales in between, are characterized by a number of phenomena such as fronts, gravity waves, thunderstorms, wind jets etc. The mesoscales tend to be defined in terms of the adjacent boundary-layer and synoptic scales, rather than in their own right (Paper IV).

Other classifications of the atmospheric spectrum of motions exist. For instance Stull (1988) distinguished between macro-, meso- and micro-scales, in descending order, while Thunis and Bornstein (1996) suggested three subdivisions of each scale to achieve a total of nine classes. There exists no clear-cut definitions of the limits between any of the scales and classes. The most poorly defined limits are those related to the mesoscale and boundary-layer scales (Paper IV). Tentative limits on the scales are given in Figure 2.1. Indeed, while the synoptic and planetary scales each span roughly one order of magnitude, the mesoscales contain three and the boundary-layer scales encompass motion over six orders of magnitude.

Each scale of motion has a certain amount of kinetic energy associated with it. Deriving an average energy spectrum from atmospheric winds reveal several things about the atmospheric system. Spectra come in two kinds. Frequency spectra are obtained from a time-series of observations at a point, while wavenumber spectra can be obtained from for example aircraft ob-

servations. The wavenumber is the number of waves per length, i.e. the inverse of the length scale indicated in Figure 2.1. An idealized energy spectrum is illustrated in this figure. It is based on knowledge obtained from large boundary-layer experiments in the 50's to 70's (Panofsky and Van der Hoven 1955; Fiedler and Panofsky 1970; Vinnichenko 1970; Stull 1988). These 'traditional' or 'textbook' spectra have peaks in the synoptic and boundary-layer scales, separated by a spectral gap in the mesoscales. The general explanation of these spectra is that two scale regimes of instability exist in the atmosphere; baroclinic instability is excited by horizontal gradients of wind and temperature on the synoptic scales, while shear- and buoyancy instabilities are excited by vertical gradients. Parts of the energy spectrum are predictable from fundamental theories, for instance, it can be shown that the tilt of the spectrum at scales smaller than the peak in the boundary layer scales should be a power law with energy proportional to the wavenumber to the power of $-5/3$ (Kolmogorov, 1941). Here, it is assumed that all the energy is generated at the peak-scale and subsequently dissipated at the viscous scale at exactly the same rate. The smaller the scale the more effective the transfer of energy is, hence the eddies contain less energy. Interestingly, the mesoscales are often observed to have the same slope ($-5/3$), while it is steeper in the upper synoptic-scale range (-3), based on sensors placed aboard commercial aircrafts (Nastrom et al. 1984).

Models of the atmosphere are usually built on grids. The grid nodes represent the mean properties, wind, pressure, temperature etc. of the atmosphere in a small box around it. The distance between gridboxes determines the model resolution, and, popularly speaking, the motion on scales larger than the grid spacing is resolved¹. Motion at scales smaller than the grid spacing is accordingly referred to as unresolved. These scales need to be parameterized by relating their effects on the resolved scales, to the resolved motions themselves. Different types of atmospheric models exist. In fact an enormous wealth of models exists, each with their own advantages and drawbacks, each applicable to a number of problems. Some cover the entire Earth, which we classify as global models, while others cover only a limited area. Global models are frequently used in both operational weather forecasts and climate studies. Limited-area models rely on global models for their boundary conditions. These models were first developed for weather forecasting, because they allow finer resolution, but have recently been used in regional climate studies (e.g. Tjernström et al. 2005). Limited-area models allow a higher resolution at the expense of global coverage.

In fact resolution is a key issue here. Computers are a limiting factor for our models, and will continue to be so in the near future. Even the most powerful computers are unable to cover more than three orders of magnitude in scale,

¹Strictly speaking only motions of two times the gridspacing are explicitly resolved. This is known as the Nyquist scale. Further, we consider motions to be well-resolved when they stretch, at least say, five to ten gridpoints.

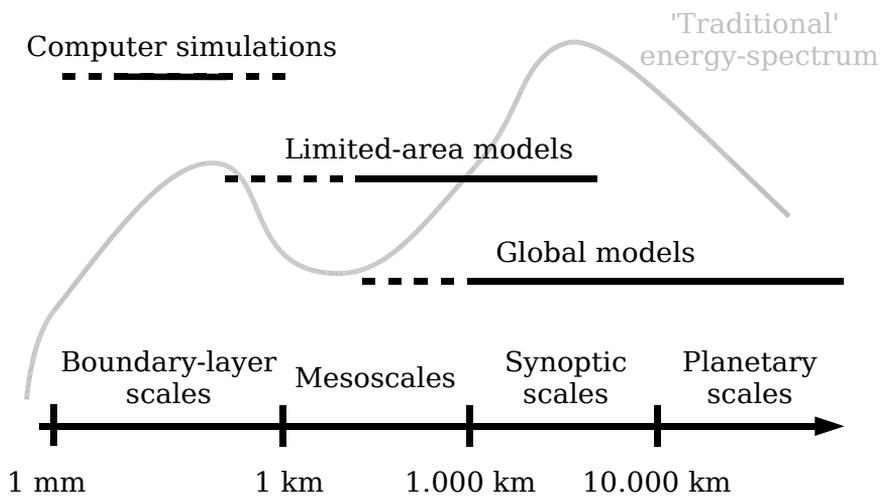


Figure 2.1: Atmospheric scales of motion. The physical limits and the shape of the energy-spectrum are tentative.

while two orders are more common (Figure 2.1). For example the European Center for Medium Range Weather Forecasts (ECMWF) global model, one of the finest models for weather prediction as of today, is able to resolve at best between one and 799 waves around the Equator. The problem will continue to exist for a long time because a doubling of the resolution results in a ten-fold increase in computational loading. Hence going from three to four orders of magnitude scale coverage requires a thousand times more computational power. The same limitations apply to fine-scale computer simulations, which are frequently used to study boundary-layer scale turbulence. For example direct numerical simulations (DNS) carry gridspacings of a few millimeters, and hence are unable to deal with a volume larger than, say, a few cubic meters (Figure 2.1). Large-eddy simulations (LES) parameterize the small-scale turbulence and consider only the largest turbulent eddies. Even so they seldom cover more than two orders of magnitude and are therefore unable to emulate interactions with the mesoscales to any significant extent. Consequently, our knowledge of the atmosphere based on computer models and simulations remains patchy. We can only simulate a small range of scales at a time, for instance the model used in Paper I spanned 1-100 km. The larger and smaller scales of motion are either handled though boundary conditions, neglected or parameterized. For example, the boundary conditions for limited area models are usually obtained from global models, extending the resolved scales of motion, while LES/DNS often use cyclic boundary conditions whereby motion over scales larger than the domain are neglected.

It was suggested by Osborne Reynolds (1848-1912) to divide turbulent flows into two parts; the mean and the turbulent deviation from the mean. Summing the two parts we retain the original flow. Formally, we write the Reynolds decomposition of a single variable as:

$$u(t) = \bar{u} + u(t)', \quad (2.1)$$

where x , y and z are spatial dimensions and t is time. The overline defines the time average of the variable and the prime the temporal deviations from the mean. It follows that $\overline{u'} = 0$. We shall further use standard notation, such that u is the wind in the x -direction, v is the wind in the y -direction and w is the vertical wind. Reynolds averaging is used for analyzing observations and in the theoretical derivation of equations for the mean-flow and the higher order moments, including velocity variance $\overline{u'u'}$ and vertical fluxes of momentum and heat, $\overline{u'w'}$ and $\overline{w'\theta'}$.

The rationale behind the use of global and limited-area models, which often only resolve the global and synoptic scales, is that the unresolved motion that is important can be described in terms of the resolved scales, aided by the Reynolds decomposition. That is, given a certain state of the resolved scales the unresolved scales will respond, at least statistically, in a predictable way. If so, we can create parameterizations, based on theory and observations, that replace the unresolved scales (e.g. Paper II and Paper III). The development of such parameterizations is aided by the presence of the spectral gap discussed above. The spectral gap allows separation of the boundary layer scales from the synoptic scales. If we account for everything from the smallest scales up to the peak of the boundary layer scales, including more scales will not add significantly to the results. In that case, the same parameterization for unresolved scales can be used in any model with grid spacings between 1 and 1.000 km, because these scales are not all that important. If, on the other hand, the spectral gap does not exist, the necessary parameterization will depend on the resolution (Duynderke, 1998).

As simple as it may seem, the Reynolds decomposition (2.1) offers great practical difficulties when applied to observations (Paper II). The source of trouble is the time-average² that has to be performed when separating the mean-flow from the deviations. The expression '*one man's turbulence is another's mean-flow*' certainly applies. If there is gap in the energy spectrum it should not matter much, as long as we have included the boundary layer scales. Spokesmen for a shorter time-window argue that the mesoscale motion should be avoided as it does not follow any laws or simple rules (Mahrt and Vickers 2003). The critics say that we need to investigate the influence

²In principle, an ensemble average over all possible realisations of the experiment has to be taken. This is most inconvenient, since the atmosphere only offers us one of these possible realisations, and in practice we therefore only consider time- or space-averages, aided by Taylor's hypothesis of 'frozen' turbulence.

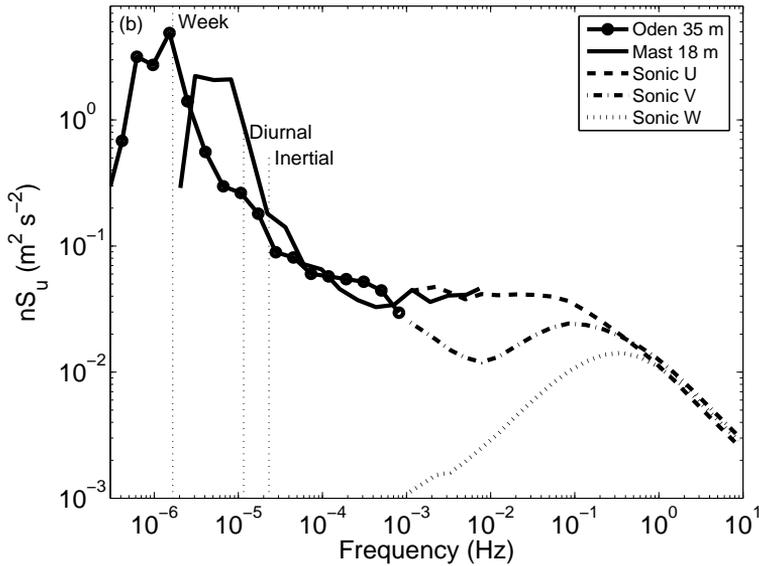


Figure 2.2: Frequency energy-spectrum obtained during the AOE-2001 campaign. The boundary-layer scales are divided into along wind, u , cross wind, v , and vertical, w , deviations from the mean. These are measured by the sonic anemometers. Note that the smallest scales are on the right, opposite to Figure 2.1. See Paper IV for further details.

of the mesoscales as they are potentially important in global models (Beljaars and Viterbo 1998; Viterbo et al. 1999; Beljaars 2001; Chapter 4).

While the traditional power spectra (Panofsky and Van der Hoven 1955; Fiedler and Panofsky 1970; Vinnichenko 1970; Stull 1988) exhibit a clear spectral gap, measurements from the Arctic summer boundary layer indicate a plateau in the mesoscale range (Figure 2.2; Paper IV). At times there is not even a plateau, and the mesoscale motion appears to enhance the surface turbulence. Further, there are indications that mixing occurs from the free atmosphere into the boundary layer during such events. The mixing events are more frequent when the temperature difference between the surface and the free atmosphere is larger, i.e. the more stable the stratification is. Two major intrusions of warm air from lower latitudes set the background for such activity. One can only speculate where the mesoscale energy comes from. Paper IV suggests gravity waves and very shallow mesoscale fronts confined to the lowest part of the atmosphere. These results provide some support for recent theories that gravity waves in the free atmosphere may influence surface turbulence if there is a continuous stratification between the boundary layer and the free atmosphere (Zilitinkevich 2002). Further, paper I presents simulations of gravity waves emerging from openings in the sea-ice. These waves are shown to cause elevated mixing of this type within the model framework. The waves are, however, standing and should be distinguished from the waves observed in Paper IV which travel past the stationary sensors.

It can certainly be debated whether the spectral gap really exists in the stably stratified boundary layer of the Arctic (e.g. Paper IV). It has been argued that the central Arctic Ocean is 'the perfect laboratory' for studies of small-scale turbulence, since there is no topography and the surface is relatively horizontally homogeneous (Grachev et al. 2005). These are basic assumptions often violated at other experimental sites. More so, they are often blamed for being the sources of unwanted or disturbing mesoscale motion. Even so Paper IV finds evidence for frequent mesoscale activity, both features resembling gravity-waves and mesoscale fronts apparently thriving under the Arctic inversion.

3. Stably Stratified Boundary Layers

In the 1970's and 80's the stably stratified boundary layer was, more or less, put aside for studies of the unstable convective boundary layer. There can be several reasons for this neglect. The unstable boundary layer is much deeper, characterized by larger fluxes of heat, momentum and pollutants, occurs comfortably during daytime, and is in general more spectacular than its stable counterpart. Further, the computer simulations at that time did not permit the study of the stable boundary layer due to the fine resolution necessary. Further, long-standing theories predicted that if the boundary layer was very stable, beyond a certain limit to be explained below, turbulence would decay (Richardson 1920; Chandrasekhar 1961; Howard 1961; Miles 1961; Miles 1986). These results are repeated in most textbooks on the subject. The theories are now being called into question by observations (Paper II) as well as from a theoretical and modeling perspective (e.g. Zilitinkevich et al. 2007; Paper III). We shall take a closer look at this debate.

When light air reside above more dense air, we say that the atmosphere is stably stratified. Most of the atmosphere is in fact stably stratified. When the opposite occurs the atmosphere is out of balance and will attempt to overturn in order to restore the balance. We refer to this situation as the atmosphere being unstable. We have already touched upon these phenomena in the introduction - here we shall proceed in a more formal fashion. Unfortunately, the above notions on the stability are not exactly true but require some refinement. Because pressure decreases with altitude the density of a parcel of air will also decrease as we move it upward, at the same time as the temperature decreases due to the expansion. For a dry atmosphere in balance the temperature will decrease by roughly 1°C per 100 m altitude, less if the atmosphere is saturated with water vapor. What determines whether the atmosphere is stable or unstable is if the density of a parcel of air is smaller or larger than its surroundings when perturbed from its initial position. This means that the dry atmosphere is stable if the temperature decreases by less than 1°C per 100 m and unstable if the gradient somewhere is steeper. For these reasons meteorologists find it convenient to define variables that are conserved under vertical motion. Many different definitions exist, here it suffices to define the potential temperature:

$$\theta = T \left(\frac{p}{p_0} \right)^{-\frac{R}{c_p}}, \quad (3.1)$$

where T is the absolute temperature measured in degrees Kelvin, p is the actual pressure, p_0 is a reference pressure, 1000 hPa, R is the gas constant and

c_p is the specific heat capacity of dry air at a constant pressure, respectively. θ is the temperature a parcel of air would achieve if brought to the reference pressure without exchange of heat with the surroundings¹.

The potential temperature definition allows us to easily distinguish between unstable and stable stratification. If θ increases with height, the stratification is stable and if the potential temperature is higher close to the ground than above, the atmosphere is unstable. If the atmosphere is stable, a parcel released out of equilibrium will oscillate up and down about the equilibrium height. The rate of the oscillation is the famous Brunt-Väisälä frequency, N , given by:

$$N^2 = \frac{g}{\theta} \cdot \frac{\partial \theta}{\partial z}, \quad (3.2)$$

where g is gravity. Note how N^2 increases as the vertical derivative of θ does: The more stably stratified the atmosphere is the faster may it oscillate. If the stratification is unstable, i.e. $\partial \theta / \partial z < 0$, N is imaginary. The physical interpretation of an imaginary N is that a parcel brought out of equilibrium in an unstable atmosphere will accelerate away from equilibrium, ultimately resulting in an overturning of the atmosphere.

Under stably stratified conditions mixing may, however, prevent the parcel from returning to its original position. The necessary energy to overcome the stable stratification can be attained from the mean-flow wind-shear. Usually, horizontal shear is negligible in the boundary layer compared to the vertical shear. Therefore, the total shear can be approximated:

$$S^2 = \left(\frac{\partial U}{\partial z} \right)^2 + \left(\frac{\partial V}{\partial z} \right)^2, \quad (3.3)$$

where U and V are the components of the mean wind. Note how S has the same unit of frequency (s^{-1}) as the Brunt-Väisälä frequency.

Figure 3.1 shows examples of potential temperature profiles. Five different classes of semi-stationary dry boundary layers is what we can imagine (Paper III). If we introduce clouds additional classes appear. The unstable boundary layer is usually the deepest of them all. Here the layer close to ground is heated by the warm surface. As a result $N^2 < 0$, so parcels of air will accelerate upward until they reach a new equilibrium in the upper inversion. The somewhat academic neutral boundary layer has a constant potential temperature and is characterized by zero heat flux and Brunt-Väisälä frequency. The conventionally neutral boundary layer typically occurs in windy situations, where strong wind-driven turbulence keeps the lower part in thermal equilibrium with the surface. The depth of the boundary layer is limited by the stability of the free

¹In the disaster movie 'Day after tomorrow' (2004) cold air from the stratosphere comes down to the surface and causes a dramatic rapid cooling of the surface. Air at 10 km height is indeed very cold, around -50°C , but the pressure is only 250 hPa. According to (3.1) the potential temperature is $+58^\circ\text{C}$, so the Earth would have been fried, rather than frozen, according to the Hollywood movie scenario.

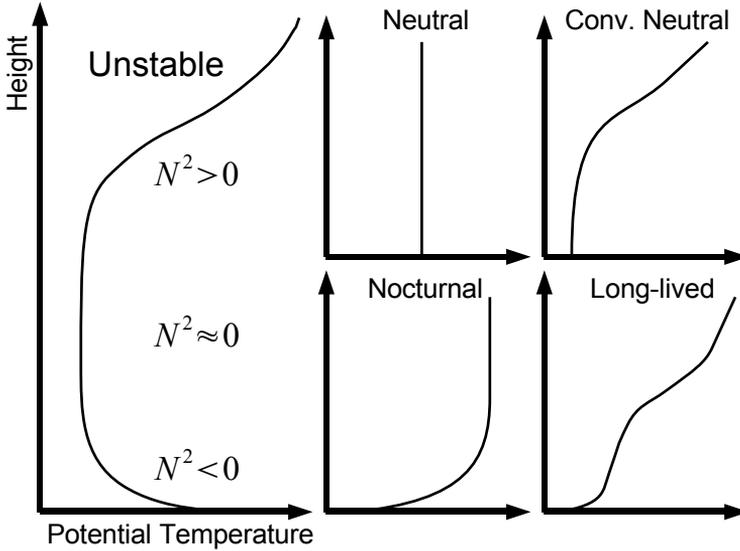


Figure 3.1: Idealized potential temperature profiles for five classes of dry unstable, neutral and stably stratified boundary layers.

atmosphere, generating the upper inversion with a large N^2 . The nocturnal stable boundary layer forms after sunset over land due to the radiative cooling of the surface. Finally, the long-lived stable boundary layer develops against stable stratification, while losing heat at the surface. These boundary layers typically occur at high latitudes and over land during winter. Note how all boundary layer classes, except the truly neutral, contain stably stratified parts. In addition most of the free atmosphere and the majority of the oceans are stably stratified, highlighting the need of a better understanding of turbulence in stably stratified conditions.

3.1 Richardson's Critical Number

Long-standing theories suggest that if the stratification, N^2 , dominates over the wind shear, S^2 , then turbulence will decay and the flow will tend to become laminar. Richardson (1920) investigated the evolution of the atmospheric motion energy of small scale disturbances, the turbulent kinetic energy (E_k). Here we shall repeat his original considerations on the problem in a modern dress. Richardson derived the budget equation for E_k :

$$\begin{array}{ccccccc} \frac{DE_k}{Dt} & = & \tau \cdot \mathbf{S} & + \frac{g}{\theta} \cdot \overline{w'\theta'} & - \epsilon & - \frac{\partial F_k}{\partial z}, & (3.4) \\ \text{Storage} & & \text{Shear} & \text{Buoyancy} & \text{Dissipation} & \text{Transport} & \end{array}$$

where D/Dt is the rate of change following the mean flow, $\tau = -(\overline{u'w'}, \overline{v'w'})$ is the stress vector, $\mathbf{S} = (\partial U/\partial z, \partial V/\partial z)$ is the mean-flow shear vector such

that $\mathbf{S} \cdot \mathbf{S} = S^2$, g is gravity, ε is the dissipation rate and F_k is the vertical transport of turbulent kinetic energy. Here we shall not consider the vertical transport term. Equation (3.4) states that the rate of change of E_k is a balance between the *shear production*, *buoyancy conversion/production* and *viscous dissipation*. The shear production is usually positive, while the buoyancy term is negative in stable- and positive in unstable stratification. Buoyancy in stable stratification is therefore a sink of E_k and is often referred to as a destructive term. We find it more appropriate to label it conversion for reasons to be given below. The dissipation term is always negative, but presumably proportional to $E_k^{3/2}$ (Kolmogorov 1941).

Next, Richardson considered a case when E_k was small, but finite, such that the shear production would not be zero, but sufficiently small that dissipation can be neglected. The question is then if the turbulence will grow, i.e. whether $DE_k/Dt > 0$? Following (3.4), this will be the case if:

$$\tau \cdot \mathbf{S} > -\frac{g}{\theta} \cdot \overline{w'\theta'}. \quad (3.5)$$

By defining the eddy conductivity $\overline{w'\theta'} = -K_h \frac{\partial \theta}{\partial z}$ and the eddy viscosity $\tau = K_m \mathbf{S}$ and rewriting (3.5) we get an inequality in non-dimensional ratios:

$$\frac{K_m}{K_h} > \frac{N^2}{S^2}. \quad (3.6)$$

The entity on the left hand side is known as the turbulent Prandtl number, $Pr \equiv K_m/K_h$, while the right hand side was later to be known as the Richardson number, $Ri \equiv N^2/S^2$, such that the inequality reads:

$$Pr > Ri. \quad (3.7)$$

Richardson assumed that $K_m = K_h$ such that the necessary condition for turbulence to grow can be stated $Ri < Ri_c = 1$, where Ri_c is the critical Richardson number. This means that if the flow stability is weak, $Ri < 1$, then turbulence will grow to a level where the production is balanced by dissipation. Contrary, if $Ri > 1$ the flow is too stable to support turbulence growth according to Richardson's results. Later theoretical studies have supported Richardson's notion by other means (Chandrasekhar 1961; Miles 1961; Howard 1961). Observations do support a Pr close to unity in near-neutral conditions, 0.7-0.8 most often being reported. However, Pr appears not to be constant, rather observations and computer simulations indicate that it is a function of Ri such that $Pr(Ri) \propto Ri$, for large Ri (e.g. Zilitinkevich et al. 2007). In this case (3.7) is always satisfied, which means that turbulence will always grow to achieve a balance with dissipation.

In practice, an increasing Pr means that momentum is being mixed more efficiently by turbulence than heat, which is supported by observations presented in Paper II. Figure 3.2 shows the momentum and heat fluxes, normalized by turbulent variances, for example τ/E_k (Paper II). The normalization

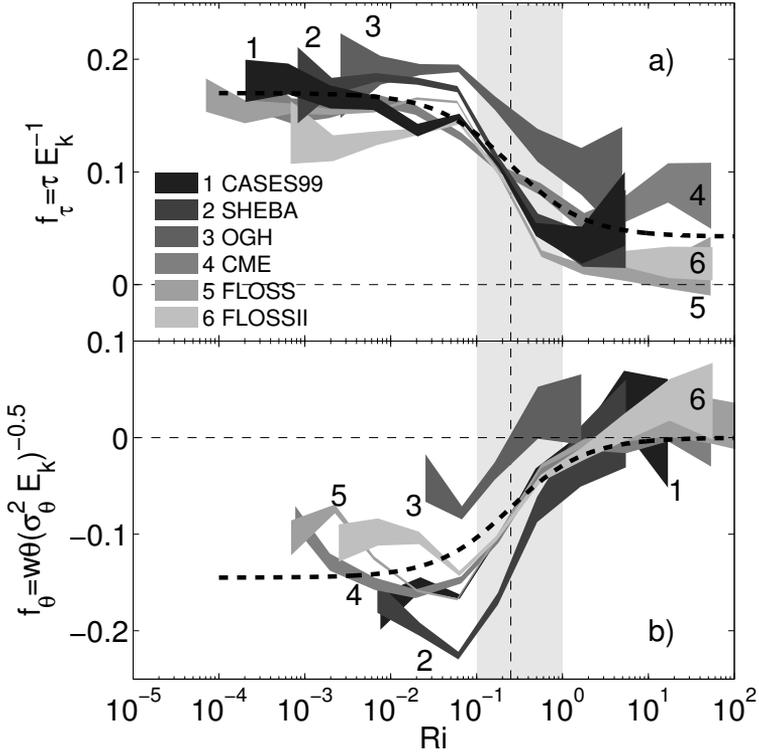


Figure 3.2: Observed normalized fluxes of a) momentum and b) heat as functions of the gradient Richardson number. Data were binned and 95 percent confidence intervals on the mean were calculated by the two-sided Student's t-test. The shapes shows the confidence intervals only. See Paper II and III for further details.

means that turbulent stresses from two observations with different levels of turbulence can be compared, the only feature linking the two being the mean-flow dynamic stability, Ri . Such analysis represents the hunt for universality in observations, despite widely varying physical conditions. Such an endeavour is based on similarity theory; a physically based connection between two non-dimensional variables. For Ri a connection of this type was established already by Klipp and Mahrt (2004) and Sorbjan (2006), so Paper II refrains from repeating it. Rather, a thorough investigation of how it works for a large amount of data is performed. A total of five and a half equivalent years worth of stably stratified data were collected from six different experiment sites.

The results indicate that there exists a range where the normalized fluxes of heat and momentum are almost constant for small Ri . We denote this range the weakly stable regime. In this regime the turbulence acts as if the flow was neutral. Turbulent eddies are so strong that they carry along parcels of air with different potential temperatures without them being able to return to their initial height. Between 0.1 and 1 there is a sharp transition, where the turbulent eddies start to feel the stability of the flow. The fluxes decrease rapidly to

achieve new nearly constant levels in the very stable regime for $Ri > 1$. Here the turbulent stress is reduced to a fraction of the weakly stable regime level, while the turbulent heat flux is statistically indistinguishable from zero. The latter does not mean that the heat flux is zero, rather we are unable to measure it such that we can have statistical confidence to conclude that it is non-zero.

The formulations above may seem somewhat cryptic to some readers. Indeed, such is fundamental to the scientific method (Popper 1934; Randall and Wielicki 1997). Hypotheses can be set forth only if they are falsifiable. It is the idea that science advances by unjustified, exaggerated guesses followed by thorough criticism. Subsequently, general consensus on their validity can occur only after thorough observational scrutiny. However, all theories and hypotheses will always be subjects of disproof. In this case, the hypothesis is that of Richardsons (1920) paper, stating that turbulence does not grow actively by itself at $Ri > 1$. It is being disproven by Figure 3.2 a) as the observations indicate a finite turbulent stress making the shear term in (3.4) positive. Figure 3.2 b) suggests, although does not prove, that this is possible because the buoyancy term is close to zero. Figure 3.3 summarizes the stability regimes suggested by the observations.

3.2 Failure of Obukhovs Length

The by far most popular means of understanding and ordering observations of stably stratified turbulence is the Monin-Obukhov (MO) similarity theory (Obukhov 1946; Monin and Obukhov 1954). Almost all atmospheric models in use today base their turbulence closure models on it, and nearly all observational and theoretical studies of stratified atmospheric turbulence and most textbooks on the subject are somehow based on, or contain elements of, MO similarity. Instead of Ri , the non-dimensional height above ground is used as the stability parameter:

$$\frac{z}{L} = \frac{gkz}{\theta} \cdot \frac{-\overline{w'\theta'}}{\tau^{3/2}}, \quad (3.8)$$

where $k \approx 0.4$ is the von Kármán constant, z the height above ground, and L the Monin-Obukhov length. Originally, z/L was intended to be applied only to the lowest part of the boundary layer, the surface layer, where the fluxes can be assumed to be approximately constant. Later the idea was extended to the interior of the boundary layer, by evaluating the heat flux and stress locally at the height under consideration (Nieuwstadt 1984). The procedures result in relations between the mean flow gradients of wind, temperature and tracers and their respective vertical fluxes. As such they can be used to predict the fluxes in models, given the mean.

Results from MO-based observational studies, much like the one shown in Figure 3.2, exhibit impressive coherence and collapse of data on nice curves and lines; something that has been accepted as token for the validity of the MO

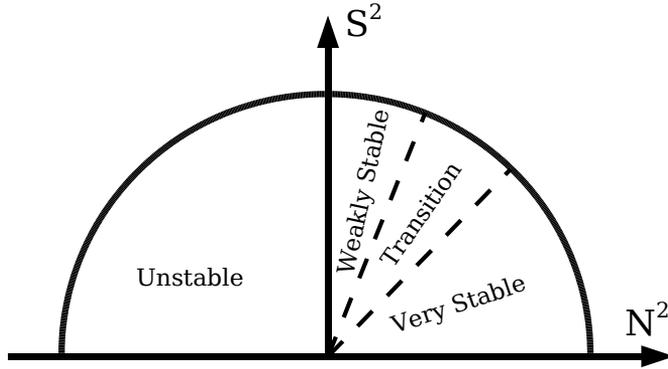


Figure 3.3: Schematic of the mean-flow stability range.

similarity theory. Further, the results have supported Richardson's hypothesis by showing fluxes that tend to zero for large z/L . Unfortunately, data analysis based on z/L is subject to self-correlation (e.g. Hicks 1978; Mahrt et al. 1998; Baas et al. 2006). Self-correlation occurs in data analysis when the same measurement is used on both axes in a plot. For example, a plot of τ/E_k versus z/L would include τ on both axes. Since every measurement has a certain random error associated with it, these errors will often correlate in such plots to generate an artificial behaviour of the datapoints. In this case, an erroneously small τ will cause the z/L estimate to be erroneously large, forcing the curve down towards zero in the very stable regime. The problem of self-correlation is a general one for studies of turbulence data and is very difficult to avoid. Papers II and III lay out a possible path to achieve this goal, though beware, it is not free from trouble and we have yet to see the end of it.

4. Models of Atmospheric Turbulence

The smallest scales of motion, the boundary layer scales, are not resolved by global and limited-area models. Since they are important for the results they need to be included by parameterization as discussed in Chapter 2. Arguably, the most important task of a turbulence parameterization is to provide the surface drag to the large-scale flow, and furthermore to provide the atmospheric exchange of heat, moisture and tracers with the surface. With the evolving demands on, and complexity of, atmospheric models, the prediction of near-surface variables such as temperature, wind and boundary-layer cloudiness are becoming increasingly important (e.g. Beljaars and Viterbo 1998). This means that correct representation of not only the surface exchange but also the vertical structure of the boundary layer is presently in focus. When designing and developing a turbulence parameterization scheme several practical problems have to be considered. Ideally, the scheme should be based on physical principles, and behave as nature does. Unfortunately, the first requirement is complicated by the turbulence closure problem; even though we know the exact equations that describe turbulent flows, we are in general unable to solve them. Further, even given a good parameterization, modeling the boundary layer is a non-linear and coupled problem, which makes the solution very sensitive to even small changes in the scheme. The scheme should function across a wide range of conditions, day and night, summer to winter, from Sahara to Siberia, and in principle on other planets. Probably, these constraints are part of the reason why we are reluctant to include new parameterizations in forecast and climate models. Most models base their parameterization on developments from the 1970s and early 80s.

The properties of the parameterization affect the usefulness of short- to medium range weather forecasts and climate simulations in different ways. For example short-range weather forecasts for 1 to 2 days are often performed to obtain details at and close to the ground, such as temperature, frost, fogs, strong winds etc., while medium-range weather forecasts for 5 to 10 days will be most concerned with the development of cyclones and anti-cyclones. In the former case a well-known problem is so-called run-away cooling. At night, when the surface cools rapidly, these models predict that all turbulence decays in such a way that nothing stops further cooling. In the case of medium-range weather forecasts, the life-time of cyclones is largely determined by a phenomenon called boundary layer pumping or Ekman-pumping (e.g. Ekman 1902; Beljaars and Viterbo 1998; Beljaars 2001; Holton 2004; Beare 2006).

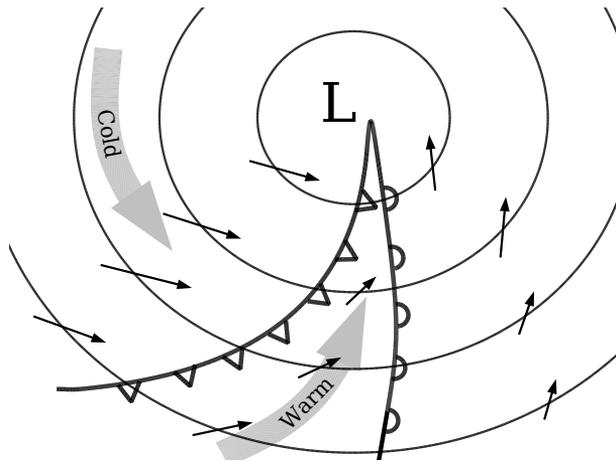


Figure 4.1: The principle of cyclone-filling by boundary layer pumping. The contours are mean sea level isobars, arrows are low-level winds and idealized cold and warm fronts and advection are shown.

Synoptic-scale flow is almost in geostrophic balance, which means that the wind is directed along the isobars. As sketched in Figure 4.1, winds would in this case circulate around the low-pressure centre. However, close to ground the turbulent drag causes the wind to weaken and turn towards the cyclone centre. This low-level convergence is balanced by an upper-level divergence. The wind-pattern causes an exchange of fast-rotating air from inside the cyclone with slow-rotating air from the outside, which further dampens the cyclone. Climate simulations are also sensitive to the representation of near-surface temperature and cyclones, though at climatic time-scales additional complications arise (Garratt 1993; Garratt et al. 1993). For example the atmosphere and ocean will have to communicate momentum and heat through the boundary layers in both the atmosphere and the ocean, hereby setting up the ocean currents and the northward heat transport. Further, the representation of the hydrological cycle depends on evaporation, which is partly controlled by boundary-layer processes.

Our understanding of and interest in the modeling of atmospheric turbulence appears to have evolved from the surface and upwards. Traditionally, the atmosphere is divided into vertical layers. The lowest few millimeters are the *viscous sublayer*, then comes the *surface layer*, the *boundary layer* and the *free atmosphere*¹. In the viscous sublayer, turbulent eddies are forced to be so small that they dissipate rapidly by molecular diffusion (Prandtl 1929). Here gradients of wind and temperature can be very large, as molecular diffusion is a relatively slow process. In the surface layer, the size of the eddies is limited by the distance to the ground. The surface layer is mainly a mathematical construct, where it is assumed that the fluxes of momentum and heat are ap-

¹The free atmosphere includes the troposphere, stratosphere, mesosphere and thermosphere.

proximately constant. This assumption permits a particularly simple solution, where the profiles of wind and potential temperature are logarithmic functions of height. The boundary layer above the surface layer assumes many forms as shown in the previous chapter, cf. Figure 3.1. The free atmosphere is thought to not be significantly influenced by turbulence; probably it is more correct to say that our knowledge of the influence is limited. The division outlined here is not unique, for example Grachev et al. (2005) named no less than six different layers, or regimes as they may be called, for the Arctic stably stratified boundary layer.

In atmospheric models this way of dividing the atmosphere is often utilized, as the temperature and winds are only calculated in vertically discrete layers. The viscous sublayer and the surface layer are assumed to exist between the surface and the first model level, regardless of the height of this level. This assumption allows relatively simple boundary conditions from a surface-layer model. Exchange between levels above the first model level is achieved via the turbulence parameterization. Some models distinguish between the boundary layer and the free atmosphere. For example, Swedish HIRLAM use one turbulent exchange model within the boundary layer and another less diffusive model in the free atmosphere. Although a division, with different models in the surface layer, boundary layer and free atmosphere, may be convenient for the programmer, there is no physical justification for it. In some cases the models do not match, and hence discontinuities of the fluxes between the surface and the boundary layer occur (Svensson and Holtslag 2007). Ideally, the same fundamental principles should apply in all three layers, and only the mathematical solution technique could possibly differ. Paper III demonstrates how a turbulence closure model can be extended to all three layers.

For mainly technical reasons, in order to allow long timesteps, the turbulent exchange within the boundary layer is described as turbulent diffusion via the eddy viscosity and eddy conductivity, K_m and K_h (see also Chapter 3). If a potential temperature gradient exists, the eddy conductivity tells us how fast turbulent motion will attempt to reduce the gradient. Physically, a strongly turbulent flow is more diffusive than a weakly turbulent flow, so the turbulent diffusivity will depend on the flow. Unfortunately, these quantities are not measurable with current observational techniques, rather they have to be calculated from measured fluxes and gradients. Models of turbulence also predict different levels of eddy viscosity and conductivity, given a fixed mean-flow. The response determines the character of the model and ultimately the state of the boundary layer it will produce. We often speak of more, or less, diffusive models. A model of the former kind will reduce the gradient of the wind between the free atmosphere and the Earth's surface where it is zero, making the boundary layer deeper than in a less diffusive model. Further, it will transport heat more efficiently towards the surface under stable stratification, making the stratification weaker and the surface warmer. Moisture will be mixed over a deeper layer, possibly obliterating low-level fogs and gener-

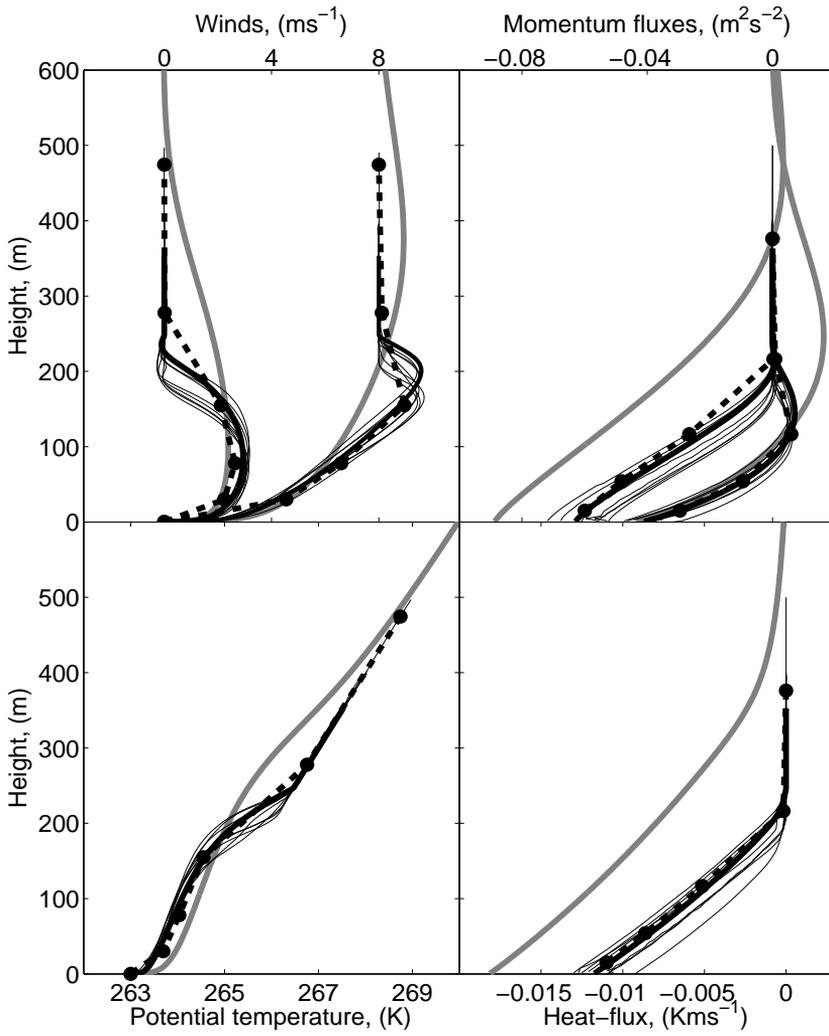


Figure 4.2: Comparison of profiles with the first GABLS case. The thick solid line is the model presented in Paper III and thin solid lines are the participating LES. The thick dashed line is a rough resolution version of the presented model, where the large dots denote the computational levels. The thick solid gray lines are for the Viterbo et al. (1999) closure.

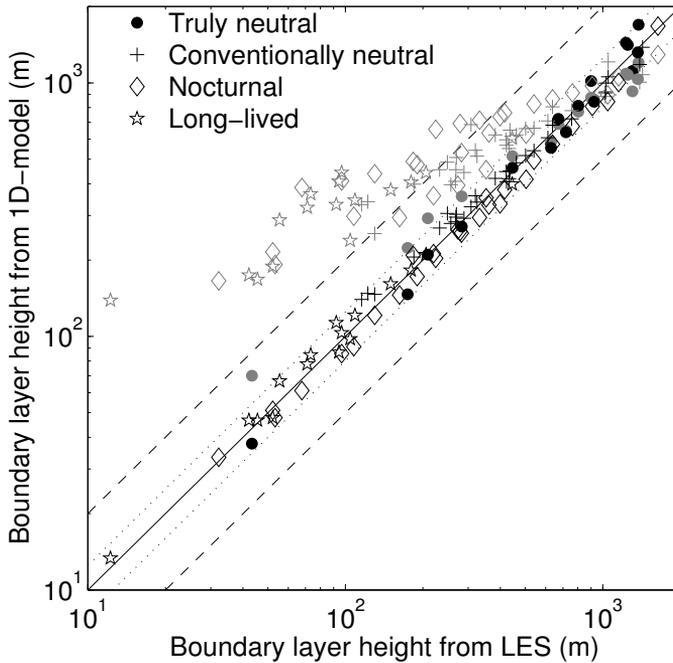


Figure 4.3: Comparison of boundary layer heights from LES and 1D-models for 90 different cases. The boundary layer classes are shown in Figure 3.1. Gray symbols are for the Viterbo et al. (1999) closure, while black symbols are for the turbulence closure model presented in Paper III. The solid line is one-to-one correspondence, while the dotted and dashed lines are 20 and 50 percent difference, respectively.

ating clouds in the upper parts of the boundary layer instead. An example of two quite different models is shown in Figure 4.2. The more diffusive turbulence closure model (Viterbo et al. 1999) produces a deeper boundary layer, with a stronger downward heat-flux, and more surface stress than the LES and Paper III model. Unfortunately, this first experiment had a fixed surface temperature, so the turbulence was not able to alter the surface temperature. Note also that the cross-isobaric massflux, i.e. the area under the left-most wind-profile, is nearly twice as large for the more diffusive model. While the first GABLS case is only considered a weakly stable case, Paper III compared model results with an LES database from the neutral to moderately stable boundary layer (Figure 4.3). The less diffusive model was able to predict the boundary layer height to within 20 percent of the LES. The more diffusive model overpredicted boundary layer height for cases shallower than 1 km and underpredicted the deepest neutral and near-neutral cases.

4.1 Boundary Layer Models

A hierarchy of turbulence closure models for the boundary layer exists. The simplest models in use are the first-order closure models, where the fluxes of heat, momentum and tracers are simple functions of the mean-flow wind and temperature. A first-order turbulence closure model parameterizes eddy viscosity and eddy conductivity under neutral and stable stratification as (e.g. Louis 1979; Viterbo et al. 1999):

$$\begin{aligned} K_m &= f_m l^2 S, \\ K_h &= f_h l^2 S, \end{aligned} \quad (4.1)$$

where f_m and f_h are non-dimensional functions, l is the turbulent mixing length and S is the magnitude of the vertical windshear. Most first-order models will have the stability functions and mixing length defined as algebraic functions. Both the stability functions and the mixing lengths are important for the behaviour of the model. Higher-order models include those solving only the turbulent kinetic energy equation (3.4), full second-order models solving nine additional equations (Mellor and Yamada 1974), and so on to infinity. The more complex the model is, the more calculations have to be done and the more parameters have to be determined. The increasing complexity of higher-order models is not necessarily beneficial for their predictive performance (e.g. Cuxart et al. 2006). The model presented in Paper III includes a prognostic equation for the total turbulent energy, E , which is the sum of the turbulent kinetic and potential energies:

$$E = E_k + E_p. \quad (4.2)$$

The turbulent potential energy, which is proportional to the temperature variance, is zero for neutral stratification, while Paper II found that at very strong stability E_p is roughly half of E_k . In contrast, the motion of a pendulum swinging back and forth also consist of both kinetic and potential mechanical energy. At the low point all the energy is kinetic, while at the extremes the energy is all potential. In the absence of friction, the sum of the two will be constant. In Chapter 3 we demonstrated that the use of the turbulent kinetic energy equation (3.4) may lead to implicit critical Richardson numbers. As demonstrated in Paper III, adding the turbulent potential energy to the kinetic actually simplifies the problem to solving:

$$\begin{array}{ccccccc} \frac{DE}{Dt} & = & \tau \cdot \mathbf{S} & -\gamma & -\frac{\partial F_E}{\partial z}, & & \\ \text{Storage} & & \text{Shear} & \text{Dissipation} & \text{Transport} & & \end{array} \quad (4.3)$$

where now γ is the total dissipation rate and F_E is the total turbulent energy transport. An additional source term appears in unstable situations. Given E , we can calculate the relative contributions from E_k and E_p and via Figure 3.2 we can calculate the fluxes of momentum and heat. Finally, the dissipation rate

is parameterized using the dissipation length-scale (Kolmogorov 1941). Employing equation (4.3) instead of (3.4) eliminates the implicit critical Richardson number, as it allows turbulence to exist as long as $\tau \cdot \mathbf{S}$ is non-zero.

In the end, almost all models we can think of contain some kind of turbulent length-scale², be it mixing-length or dissipation length. The turbulent length-scale is our way of obtaining closure and it needs to be specified somehow. At a philosophical level the turbulent length-scale reflects our current understanding of the problem at hand, such that its definition leaves us with considerable artistic freedom. In more practical terms the length-scale determines the amount of mixing, according to (4.1). The larger the turbulent length-scale is, the more mixing is produced by the closure. Almost every paper on the subject finds its own flavour of the turbulent length-scale. General consensus exists that the length-scale close to ground is limited by the distance to the ground such that:

$$l_z \approx kz, \quad (4.4)$$

where $k \approx 0.4$ is von Karman's constant. As we move away from the surface the turbulent eddies no longer feel the presence of the surface, a region known as the *z-less* layer. Here the turbulent eddies are limited in size by other processes. These may vary with height, though they are not explicitly dependent on height. A common way to parameterize the length-scale for the truly neutral boundary layer was suggested by Blackadar (1962):

$$l^{-1} \approx l_z^{-1} + l_o^{-1}, \quad (4.5)$$

where l_o is the asymptotic length which is usually set to some fixed value, for example 150 m is often used. Close to the ground l will behave as kz and far from the ground it will approach the constant value of l_o . Observations indicate lower values, for example Tjernström (1993) found 23 m. However, there is no physical reason that a constant value of the length-scale should apply to all possible situations in the atmosphere. Rather, it has been suggested (e.g. Zilitinkevich and Mironov 1996; Paper III) to exchange l_o with two physically-based limits. The length-scale for neutral stratification (Rossby and Montgomery 1935; Zilitinkevich 1972):

$$l_f \approx C_f \frac{\sqrt{\tau}}{f}, \quad (4.6)$$

and for the influence of stable stratification (Pollard et al. 1973):

$$l_N \approx C_N \frac{\sqrt{\tau}}{N}, \quad (4.7)$$

where C_f and C_N are dimensionless constants. The length-scales can conveniently be combined as a sum of inverses $l^{-1} \approx l_z^{-1} + l_f^{-1} + l_N^{-1}$, just like

²For example, some models use a time-scale instead, though the basic idea is the same. Other models attempt to predict it, though apparently with little success for stable stratification.

the Blackadar length-scale. Larger turbulent stress yields longer length-scales, while the Coriolis force and the stable stratification limits the size of the turbulent eddies. Reasonable intervals for the constants C_f and C_N were determined in Paper III.

The beauty of using these length-scales, (4.6) and (4.7), is that they are physically-based and that they are local. In some implementations of similar length-scales, the surface friction, u_* , is used instead of $\sqrt{\tau}$. This is probably good enough for idealized cases, however, real situations are often so complicated that there can be a turbulent boundary layer and an elevated layer of turbulence in the same area. The elevated layer of turbulence may be generated by a breaking gravity wave (Paper I) or by windshear in for example a low-level jet, and as such, it will not know about the surface stress. Therefore, it would seem more plausible to use a completely local formulation for the turbulent length-scale.

4.2 More – Or Less – Diffusion

Leading forecast centers, such as the ECMWF, use turbulence closure models that deliberately are too diffusive in stable stratification (e.g. Louis 1979; Viterbo et al. 1999; Beljaars 2001). As a result the boundary layer is too deep (Figure 4.3), the near-surface ageostrophic wind-angle too small (Paper III), and the boundary layer wind maximum is placed too high compared to observations (Anton Beljaars, personal communication). We are all well aware of these deficiencies, but still we are cautious about implementing schemes that perform better in these respects. As outlined in the beginning of this chapter, the turbulence closure in an atmospheric model serves several purposes and interacts with many other schemes that account for soil processes, clouds, radiation etc. The outcome of the forecast is not only dependent on changes made in one of these schemes, it is the collective response of all of these schemes to the small change that matters.

A classic example is the run-away cooling that may occur at night or during winter in models that are less diffusive. The problem arises due to an interaction between the cooling surface and the turbulence closure scheme. When the surface starts to cool the stratification close to ground increases. Turbulence is then diminished, or disappears in models that obey a critical Richardson number, leading to a reduced downward heatflux and further cooling of the ground. A simple cure is to use a turbulence closure model that is more diffusive in order to hamper the run-away cooling (Viterbo et al. 1999). The other prominent problem is the development of cyclones. As mentioned earlier, the decay of mid-latitude cyclones is to a large extent controlled by the drag at the land-surface. The more diffusive the turbulence closure scheme is, the more drag does it produce and the stronger is the Ekman pumping. Now it appears, at least in the ECMWF modeling framework, that the cyclones live too long if

a less diffusive scheme is used. Again, the problem is solved by using a more diffusive scheme, leading to a marked improvement in the medium-range forecasting skill (Beljaars 2001).

One may speculate why the general circulation models appear to favour more mixing under stable stratification than what can be justified by observations. Global-scale models deal, in principle, with large boxes of 10-100 km in the horizontal. As explained in Chapter 2, this means that the mesoscales are not resolved, and these could potentially be part of the explanation. For example, it follows from the triangle inequality that the mean wind averaged vectorially over a box will be smaller in general than the mean windspeed averaged over the same box. The difference is controlled by the amount of mesoscale activity. Therefore, the surface fluxes of heat and momentum may be underestimated if we use the box average wind. However, it cannot be ruled out that the real problem with run-away cooling instead lies in the soil or radiation parameterizations (e.g. Steeneveld et al. 2006a; 2007). Further, suspicion at the moment points to the parameterization of the convective boundary layer, as it appears that models do not produce enough drag when the flow is statically unstable. Stable stratification occurs in the cyclone warm-sector, while convectively unstable conditions prevail in the cold advection behind the low (Figure 4.1). It can be speculated that the improvement in predictive skill obtained by increasing drag in stable stratification might have compensated for the lack of drag in unstable stratification (Anton Beljaars, personal communication). The problem as it stands, however, remains unsolved.

5. Climate and Arctic Amplification

Arctic average surface temperatures have increased by about 1.5°C during the twentieth century, which is twice as much as the global average increase of 0.7-0.8°C (Figures 1.2; Houghton et al. 2001; ACIA 2005). The phenomenon that the Arctic temperature changes more rapidly than that of the globe, both up and down, has been coined polar amplification or the Arctic amplification. The ongoing and projected climate changes have received much attention in the scientific community and in the media, with new alarming reports nearly on a daily basis. The increasing temperature has pronounced effects on the Arctic nature, wildlife and inhabitants. For example, the sea-ice extent has decreased by 10-15 percent since 1979 when satellites became available, glaciers and tundra are thawing, global sea-level is rising and vegetation is changing. Most of these effects can be attributed to the increasing Arctic temperatures.

An often quoted hypothesis for explaining the Arctic amplification is the surface albedo feedback mechanism (e.g. Manabe et al. 1991; Manabe et al. 1992; Hansen et al. 1997; Houghton et al. 2001; ACIA 2005; Bony et al. 2006; Serreze and Francis 2006). Snow and ice function as highly reflective surfaces for incoming visible sunlight. When the surface temperature increases, some of the snow and ice may melt, making the surface less reflective. Consequently, more sunlight is absorbed rather than reflected at the surface and the surface warms even further, aiding the melting even more.

Such chains of two or more physical processes either amplifying or dampening a small temperature change are named positive and negative feedback mechanisms, respectively. The primary mechanism that keeps the Earth's climate in check is the negative Planck feedback. It is due to the fact that all bodies with temperature higher than the absolute zero radiate energy proportional to the absolute temperature to the fourth power. This is the case for the Earth, where the effective temperature is determined by a balance between the absorbed sunlight and the outgoing Planck radiation. The emitted radiation is in the infrared due to the low temperature of Earth. An increase in the effective temperature would result in an imbalance that would attempt to restore the lower equilibrium temperature. It is important to distinguish internal feedback mechanisms from external forcings. External forcings include changes in the Earth's orbit around the Sun, changes in solar activity, anthropogenic emissions of greenhouse gases and aerosols that alter the radiation balance, volcanic eruptions etc.

Globally important, apparently positive, feedback mechanisms include the surface albedo, water vapor and cloud feedbacks (Bony et al. 2006). Water vapor is a greenhouse gas itself, though changes in its concentration must be considered to be of system-internal character. The amount of water vapor the atmosphere is capable of holding depends very much on the temperature. Therefore, a small warming will allow an increased water vapor loading in the atmosphere, causing a further warming. The total effect of clouds is more uncertain. At the same time as they cause a greenhouse warming by absorbing infrared radiation, they also reflect some of the incoming solar radiation back to space. The net effect depends on cloud-type and region. Further, it may be very difficult to distinguish cloud effects from water vapor effects in observations, as they are intricately linked (e.g. Intrieri et al. 2002).

On the negative side, the Planck-feedback outlined above and the so-called lapse-rate feedback mechanisms dominate. The lapse-rate is the temperature change with height and is to a large extent controlled by the tropical deep convective systems. In the Tropics a small warming of the surface will effectively be mixed throughout the troposphere by cumulonimbus clouds; in fact the temperature response will be largest in the middle- to upper troposphere owing to the release of latent heat. The infrared radiation released from Earth depends on the vertical distribution of temperature, such that a temperature change in the upper parts of the atmosphere affects the balance more than a surface-temperature change. Therefore the lapse-rate feedback is negative as a small warming will result in a relatively large release of infrared radiation.

5.1 A Simple Climate Model

It is possible to formalize the concept of feedback mechanisms using a simple linearization around the present climate state, the underlying assumption being that the changes are relatively small (e.g. Bony et al. 2006). If we consider a climate system with an unperturbed equilibrium average surface temperature, T_s , that is perturbed by some external forcing, it will respond by generating an imbalance in the radiation budget by:

$$\Delta R = \Delta Q + \lambda \Delta T_s, \quad (5.1)$$

where ΔR is the radiative disequilibrium at the top of the atmosphere, ΔQ is an external forcing originating e.g. from greenhouse gases or reflective sulphate aerosols, ΔT_s is the change in surface temperature (not effective temperature) and λ is the feedback parameter; if λ is positive the climate system is linearly unstable in its present state. At equilibrium $\Delta R = 0$, so if $\lambda < 0$ we can estimate the perturbed equilibrium temperature change:

$$\Delta T_s = -\frac{\Delta Q}{\lambda}. \quad (5.2)$$

The feedback parameter can be decomposed into contributions from different mechanisms as a simple sum, $\lambda = \lambda_P + \lambda_{WV} + \lambda_C + \lambda_A + \lambda_{LR}$, where λ_P is the gray-body Planck feedback for the Earth, λ_{WV} is the water vapor feedback, λ_C is the cloud feedback, λ_A is the surface albedo feedback and λ_{LR} is the lapse-rate feedback parameter. For a black-body planet, without an atmosphere, $\lambda = \lambda_P \approx -3.8 \text{ Wm}^{-2} \text{ K}^{-1}$. However, the Earth is a gray-body and, hence, not as effective in radiating giving $\lambda_P \approx -3.2 \text{ Wm}^{-2} \text{ K}^{-1}$. The sum of feedback parameters is more well-determined than the components, such that $\lambda_{WV} + \lambda_C + \lambda_A + \lambda_{LR}$ is between $+1.5$ and $+2.6 \text{ Wm}^{-2} \text{ K}^{-1}$, which should be added to λ_P (Bony et al. 2006).

With the simple 'climate-model' (5.1) we can constrain the observed climate change to the estimated anthropogenic forcing. The third IPCC report states that the radiative forcing of only anthropogenic greenhouse gases and ozone for 2000 compared to pre-industrial times is about 2.7 Wm^{-2} (Houghton et al. 2001). Alone, this forcing would result in an equilibrium temperature change of 1.6 to 4.5°C depending on the choice of λ , i.e. much more than the observed change of 0.7 - 0.8°C . The black-body Earth temperature change is 0.7°C . There can be several explanations contributing to the disparity between estimated and observed temperature change. The climate system has not yet reached equilibrium, several negative external forcings are at play and, finally, the estimated λ could be wrong. For instance, Anderson et al. (2003) found that the cooling effect of anthropogenic sulphate aerosols may offset at least about 1 Wm^{-2} of the greenhouse gas warming, possibly more. It is most worrisome, since the future will bring cleaner technology, which can rapidly bring down the atmospheric sulphate aerosol loading, while the greenhouse gases may stay in the atmosphere for centuries. Further, vulcanism has caused a pronounced cooling of more than 0.5 Wm^{-2} in the period 1960 to 2000, while almost absent between 1920 and 1960 (Houghton et al. 2001). Taken together, this brings the observed climate change within the uncertainty range. The large uncertainty underlines the importance of increasing our understanding and modeling of the underlying physical processes.

The formalization (5.1) is only valid for the global mean surface temperature and the top-of-the-atmosphere radiation budget. When we consider a region or a part of the atmosphere, such as the Arctic or the boundary layer, the assumptions break down. Then, the atmosphere and oceans may transport heat from one place to another, altering the local energy balance (Figure 1.3, Chapter 1). It is therefore not entirely correct to speak about climate feedback mechanisms in the strict sense, (5.1), when discussing Arctic amplification. On the other hand, the fingerprint of most feedback mechanisms are most pronounced when and where the underlying physical processes act. For instance, the large subtropical marine stratocumulus regions as the one off the westcoast of California, act as a sink of heat for the global climate system as they reflect sunlight efficiently back to space. At the same time these regions

are also colder, than what they would have been had the clouds not been there. Likewise, the globally positive surface albedo feedback is only active in regions where snow and ice is present. So we expect its impact to be the largest in these regions, though it may be felt globally to some extent through changes in the atmospheric and oceanic energy transports.

5.2 Arctic Amplification

The ACIA report (2005) mentions five mechanisms, shown in Figure 5.1 that may be responsible for Arctic amplification. Some of these may be of little importance to the global energy balance, while important locally. First, as mentioned above the surface albedo feedback mechanism acts primarily in the Arctic region, hence its effect is expected to be largest here. Second, the latent heat released as a result of warming the surface is relatively small. At lower latitudes a large fraction of the surface heat budget goes to evaporating moisture from the surface to the atmosphere, while at the low Arctic temperatures the atmosphere cannot hold much vapor. Third, the Arctic boundary layer is shallow, typically a few hundred meters or less, and dynamically isolated from the free atmosphere by the Arctic inversion discussed earlier. Hence it will be easier to warm the surface in the Arctic than in the Tropics, where the entire troposphere will warm. Fourth, a reduced sea-ice extent will allow more heat to be transferred from the ocean to the atmosphere. Finally, changes in the meridional heat transport in both the atmosphere and oceans may alter the Arctic heat budget (see Figure 1.3). The ACIA list obviously does not exclude alternative explanations, as we know so little about the Arctic climate system. Not mentioned are, for instance, the role of clouds and the temperature structure of the free atmosphere.

Figure 5.2 shows, in addition to the global mean temperature, the mean temperature north of 65°N for summer and winter, respectively, between 1850 and 2006. Also shown are individual years. Autumn and spring behaves much like the winter temperature signal. The historic Arctic temperature signal confirms the presence of Arctic amplification. However, the difference between summer and winter temperature is striking; Arctic winter has seen the by far largest year-to-year and climatic time-scale variations. Further, Arctic winter has warmed more than summer overall throughout the period in general, and during the warming period 1920-60 in particular. The latter period coincides with the low volcanic-activity period (Houghton et al. 2001). During winter the sun is well below horizon most of the time, so the surface albedo feedback is not active. So, how can it be that Arctic winter is warming faster than summer?

Authors have gone to great lengths to explain the disparity between the proposed surface albedo and the seasonal signature of Arctic amplification. The general outline of arguments is that the extra absorbed heat during summer

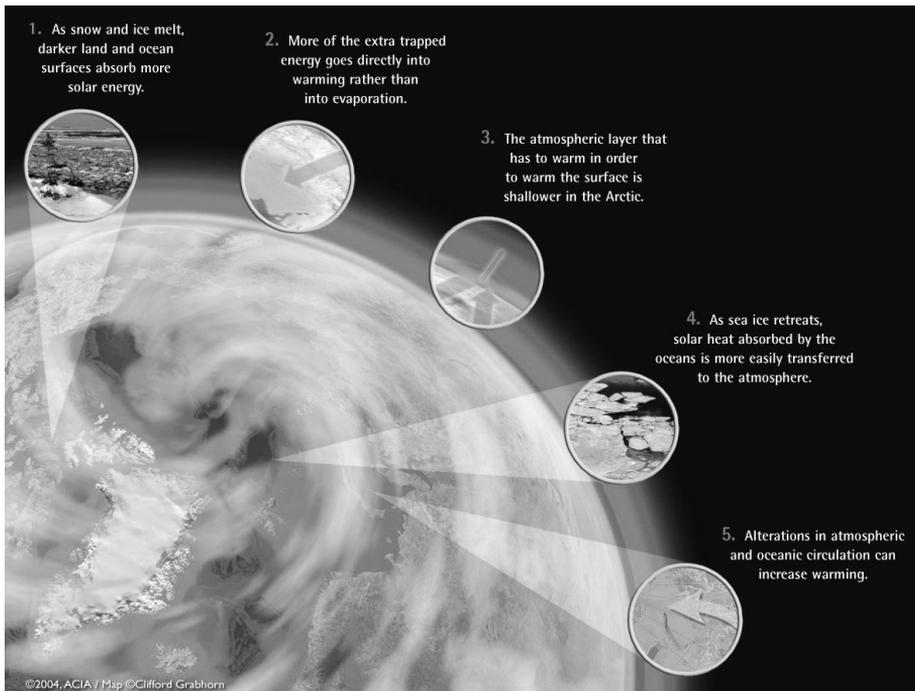


Figure 5.1: Figure courtesy granted by Paul Grabhorn, ACIA.

is used primarily for melting of sea-ice. Sea-ice has an insulating effect on the heat exchange between the ocean and the overlying atmosphere. Thus, the following autumn and winter the atmosphere will receive more heat from the ocean if less ice was present and hence be warmer. The problem with this argument is that data behind Figure 5.2 were collected from land-stations only, so no sea-ice was present. Further, the geographical distribution of the temperature signal shows the largest warming well inland in Siberia, Alaska and Canada (ACIA, 2005).

Recent idealized climate simulations indicate polar amplification even when the surface albedo feedback is artificially removed (Alexeev et al. 2005). Here simplifications of a global model were employed, including, among other things, removal of sea-ice and keeping the surface albedo the same everywhere and constant in time. The results of doubling the atmospheric CO₂ content exhibited polar amplification of the temperature increase roughly by a factor of two compared to the tropical increase. Though 'only' a model, the results suggest that the physical processes sufficient for Arctic amplification are internal to the atmosphere, for example related to atmospheric transport, water vapor, clouds or changes in the vertical temperature distribution. Obviously, the surface albedo feedback may well contribute to Arctic amplification, though it is not necessary.

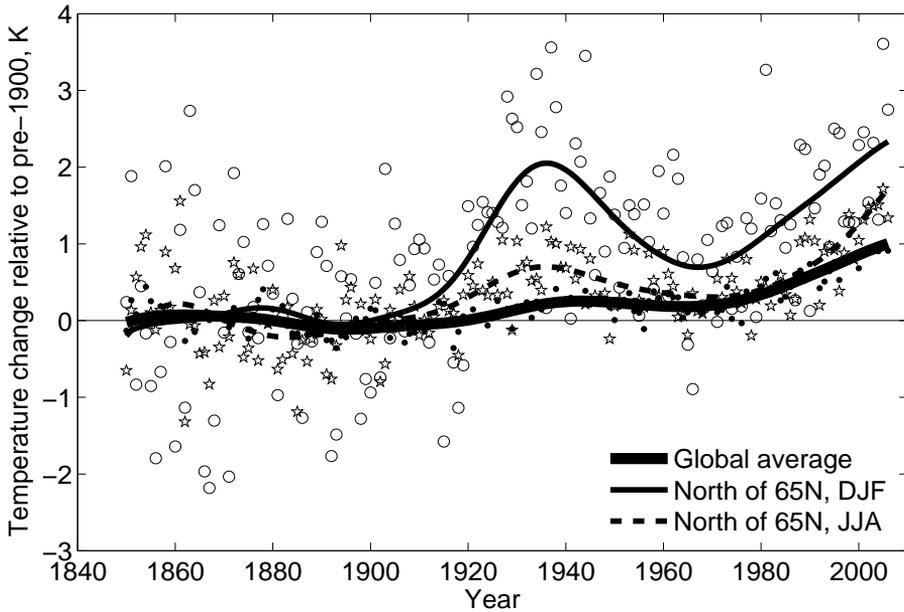


Figure 5.2: Observed area averaged relative temperature change. Only land stations are included. The symbols show the temperature in individual years, while the curves are filtered from decadal variability faster than 20 years. Dots are global, stars are summer Arctic and circles are winter Arctic. Data was provided by the Climatic Research Unit (CRU) as a 5° by 5° gridded dataset (Jones et al. 1999). The curves are offset by the pre-1900 mean value.

Picking up the thread of the third suggestion in the ACIA list of candidates (Figure 5.1), Paper V investigates the sensitivity of the dry stably stratified boundary layer to a hypothetical external surface forcing. The point that the Arctic boundary layer is more shallow than elsewhere, and hence easier to warm, can be interpreted in at least two ways. First, the thinner the boundary layer, the less heat is needed to warm it because the heat capacity is smaller. Second, the thinner the boundary layer the smaller is the layer that is able to contribute to radiating to space after a warming. The latter point can be interpreted as a kind of locally positive lapse-rate feedback mechanism, though not entirely correct, as no dynamical process analogous to the tropical deep convection acts to keep the Arctic lapse-rate fixed.

The first part, that the small heat capacity of the Arctic stable boundary layer makes it more sensitive to external forcing, is investigated in Paper V. A simple one-dimensional model setup is employed, coupling the lower atmosphere to a snow and ice layer via the surface heat balance. The turbulence closure model described in Paper III is used in the atmosphere, while the heat conduction from the snow and ice is accomplished by simple molecular diffusion. First, a long-lived boundary layer, Figures 3.1, is generated by cooling a stably stratified background atmosphere at the surface. The stability of the

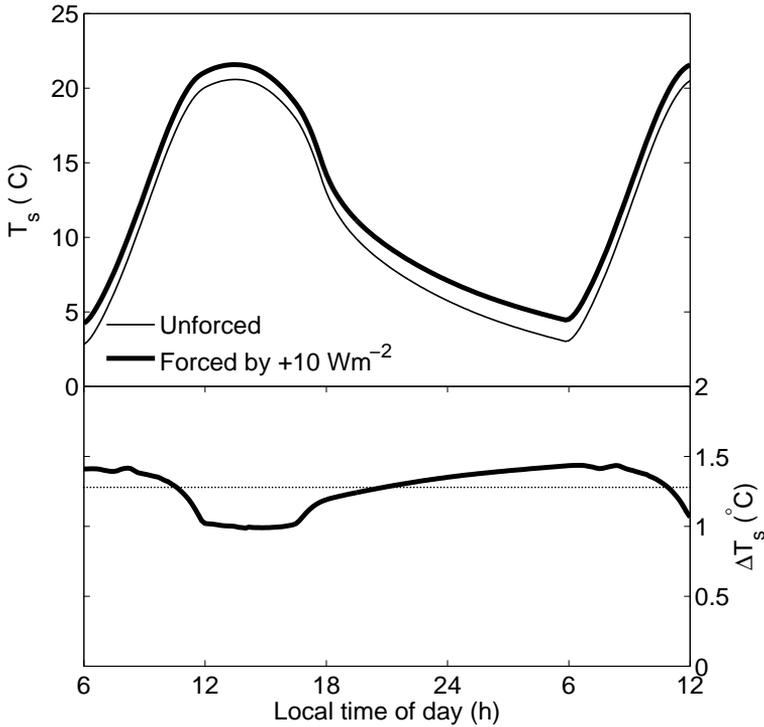


Figure 5.3: Upper panel shows modelled diurnal cycle of surface temperature for the unforced and forced cases. The lower panel is the difference between the surface temperature in the two cases. The thin dashed line is the mean change for this particular day.

boundary layer is controlled by imposing different geostrophic winds. Parallel experiments were performed by forcing the surface externally by adding $+10 \text{ Wm}^{-2}$. The results indicate that the more stably stratified the boundary layer is (i.e. the thinner) the more sensitive it is to the imposed external forcing.

Probably more illustrative and comprehensible to a general public is the simulation of the diurnal cycle of the mid-latitude surface temperature, Figure 5.3. During daytime the boundary layer is warmed from the surface by the sun. A deep unstable boundary layer develops as already discussed. During night the radiative heat loss forces the boundary layer into a stable, much more shallow, state. According to the results described above, the night-time surface temperature should be more sensitive to an external forcing than the daytime. This prediction is confirmed by the model results shown in Figure 5.3. The Arctic region has a more frequent occurrence of shallow stable boundary layers than the lower latitudes. Accordingly, we may expect that part of the Arctic amplification may be explained by their presence.

In summary, the Arctic climate system is complicated by a number of physical processes unfamiliar to, or driven into a different regime than common at the mid-latitudes and Tropics, as well as the meridional exchange of heat and

moisture with lower latitudes through atmospheric and oceanic transport. A series of mechanisms have been suggested as responsible for the Arctic amplification, though the surface albedo feedback is most frequently quoted in the literature. It is, however, being questioned if it is the driving mechanism or whether Arctic surface temperature change would be amplified even in its absence (Alexeev et al. 2005; Paper V).

6. Outlook

Understanding the Arctic climate system is one of the major problems facing climate scientists of today. The widely quoted surface albedo climate feedback mechanism blamed for Arctic amplification cannot explain the signature of the observed warming. The warming is largest over land during the polar night, when the surface albedo feedback is inactive. The debate on why the Arctic is warming more rapidly than the rest of the globe is open. Several alternative explanations exist. Paper V contributes to the debate by suggesting that the most extreme surface warming during winter over land could be due to the shallow nature of the very stable boundary layer making it extra sensitive to external forcing. Future research should attempt to evaluate the relative importance of the many processes possibly contributing to Arctic amplification.

Observations presented in Paper II question the long-standing hypothesis that turbulence should decay at large Richardson numbers. Turbulence under highly stable conditions is, however, very different from turbulence at weak stability. It may not even qualify as being turbulence in the traditional sense. The only feature that really reveals its presence is the non-zero stress, τ . In other words, a strongly stably-stratified flow will feel that an internal viscosity attempts to reduce the vertical shear. Further, the stress exerted by the turbulent motions serves to uphold the turbulence itself. Presently, we do not know the physical mechanism responsible for the presence of turbulence at strong stability. It would, however, not appear implausible that the turbulent motion at large Richardson numbers is somehow connected to internal gravity waves.

The exchange between the atmosphere, ocean and land surfaces acts through the boundary layer. This plays a key role in forecast and climate models for the representation of phenomena ranging from processes in soil, snow and ice, over the diurnal cycle and the development of cyclones, to the nature of the general circulation and the hydrological cycle. Some of the major biases present in models today can be attributed to the poor representation of the boundary layer. Any improvements we wish to introduce in these models, however, have to face the full range of phenomena that depend on the boundary layer.

There is a strong need for better communication between the scientific blocks of experimentalists, modelers and theoreticians, that are tradition within boundary-layer meteorology. An example is the improper modeling use of Monin-Obukhov similarity, which was originally only intended for

near-neutral conditions. Nevertheless, modelers have extended it to both very convective and stably stratified conditions, possibly in the lack of better? However, we have known for at least thirty years that observational data treated in this way suffer from self-correlation (Hicks 1978). One may speculate that this is part of the reason that leading forecast centers have abandoned Monin-Obukhov similarity for schemes that simply yields better forecasts for more-or-less unknown reasons. The ultimate test for the turbulence closure model presented in Paper III will be in the planned implementation in weather forecasts and climate models.

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