Long-term lidar observations of gravity waves over northern Sweden

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Abstract

This Master thesis examines observations of gravity waves with the Esrange lidar (68° N, 21° E). Wintertime measurements during the last 18 years are analyzed with a newly developed method for retrieving gravity wave parameters between 30 and 65 km altitude. The resulting ca. 1500 h of data form the foundation of the study presented in this thesis. Spectral analysis shows uniformly distributed spectra of dominant waves between 2 and 13 km vertical wavelength. Evaluating the seasonal and the inter-annual change in gravity wave activity, high gravity wave activity is found during November and December. Between January and March the gravity wave activity is reduced, which is linked to the presence of the polar vortex. Also a strong correlation between dissipating gravity waves and the monthly mean stratopause height is found. Comparison of the mean gravity wave activity measured by the Esrange lidar to earlier published studies reveals that the mean gravity wave activity in wintertime is higher at Esrange than at other stations. This is explained by the location downstream of the Scandinavian mountain ridge. The occurrence frequency of gravity wave activity shows a bimodal distribution which is linked to two different processes affecting gravity wave activity at Esrange. Furthermore, stratopause characteristics measured by the lidar are compared to those obtained from ECMWF analysis, revealing a significantly colder stratopause in ECMWF analysis. Additionally preliminary results from the Gravity Wave Life Cycle campaign are shown. These results depict gravity waves with distinct vertical wavelengths propagating from the tropopause into the lower mesosphere.
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Chapter 1

Introduction

The middle atmosphere extends from the tropopause to the lower thermosphere. This is often defined as the altitude range between 10 and 100 km. It is host to a large variety of interesting phenomena such as polar stratospheric clouds [McCormick et al., 1982], sudden stratospheric warmings [Shepherd, 2000], noctilucent clouds [Thomas, 1991], polar mesospheric summer echoes [Rapp and Lübken, 2004] and very low temperatures at the polar summer mesopause [Lübken and v. Zahn, 1991]. While there exists a general understanding of most middle atmospheric phenomena there are still unanswered questions. Prominent examples are the detailed optical properties of different types of polar stratospheric clouds [Achtert and Tesche, 2014], confirmation of the theoretical formation processes of polar mesospheric summer echoes by experiments [Rapp and Lübken, 2004] and linking observations of middle atmospheric gravity waves to their sources [Fritts and Alexander, 2003].

These incertitudes arise due to the limited accessibility of the middle atmosphere with in-situ instruments. For example balloon based measurements are only possible from the troposphere to the mid-stratosphere. Hence rockets are needed to perform in-situ measurements at higher altitudes. An alternative way to study the middle atmosphere in a continuous way is by ground-based and spaceborne remote sensing instruments. Ground-based instruments provide measurements with a good temporal resolution at specific locations. Spaceborne sensors on the other hand are capable of a higher spatial coverage. Therefore a combination of both sensor types is needed to get a more comprehensive picture of middle atmospheric processes.

It is of particular interest to study the middle atmosphere since long-term temperature trends appear to be stronger by approximately a factor of 10 in this altitude region than in the troposphere [Berger and Lübken, 2011]. Berger and Lübken [2011]
showed temperature trends of $-2$ to $-4$ K/decade at 70 km altitude between 1961 and 2009 which they linked to changes in $O_3$ and $CO_2$ concentrations. This makes the middle atmosphere a more sensitive indicator for the effects of climate change than the troposphere.

The primary dynamical coupling processes between the troposphere and the middle atmosphere are planetary waves and gravity waves [Fritts and Alexander, 2003]. These waves are mainly excited in the troposphere and transport energy as well as momentum upwards into the middle atmosphere. Prominent sources of gravity waves are for example excitation by topography, convection or shears.

Atmospheric waves are responsible for sudden stratospheric warmings in the arctic winter stratosphere [Shepherd, 2000]. These warmings result in a less efficient ozone depletion in the northern hemisphere compared to the southern hemisphere. Dissipating gravity waves, which drive an inter-hemispheric circulation in the mesosphere, are responsible for the formation of the very cold summer mesopause [Holton and Alexander, 2000]. The wave-induced acceleration that drives this circulation was estimated by Lindzen [1981] to be directed westward with a magnitude of $10^2$ m/s/day at 50 km altitude in wintertime. This provides a striking example for the importance of waves for middle atmospheric processes.

While it is well known that gravity waves have a strong impact on the middle atmosphere, they are still crudely represented in most of the large scale models, such as climate models [Fritts and Alexander, 2003]. As source processes happen mostly on mesoscale dimensions gravity waves cannot be properly resolved by large scale models, resulting in a need for parametrizations. Parametrizations often have several shortcomings: Some use non-orographic wave sources [e.g. Hines, 1997], while others are single column models which assume only vertical propagation of gravity waves, but no horizontal propagation [e.g. Medvedev and Klaassen, 2000]. Thus large scale models cannot give an accurate estimate of gravity wave forcing since source mechanisms as well as the propagation of gravity waves are not well represented [Fritts and Alexander, 2003]. Therefore a better parametrization of gravity waves is desired. An improved representation of gravity waves in large scale models can for example lead to an improved forecast of stratospheric warmings which can influence tropospheric weather conditions. Thus a more accurate weather forecast could become possible [Mitchell et al., 2013]. Also the relationship between gravity waves and climate change could be better resolved by global climate models. For example a change in gravity wave activity will result in different stratospheric winds and temperatures. This is expected
to affect surface climate and climate variability as well [Baldwin et al., 2007]. However, there are only a few long-term measurements of gravity wave activity available at the moment. To our knowledge, the longest record of gravity wave activity measured by lidars was presented by Rauthe et al. [2008] who analyzed 5 years of data. Thus it becomes difficult to estimate the long-term effect of a change in gravity wave activity.

In order to develop a better parametrization of gravity waves for large scale models, modeling studies have to be compared to actual measurements of gravity wave activity. For this an accurate climatology of gravity wave activity based on long-time measurements is needed.

The dissipation of gravity waves influences the mean circulation in the middle atmosphere. Therefore it is of particular interest to investigate in which altitude regions what amount of gravity wave energy is dissipated. Also the seasonal behavior of gravity wave activity has to be investigated as well as the geographically induced differences between different measuring stations. These are all issues that will be addressed in this thesis, since they can be addressed by lidar measurements. Lidars offer the capability to conduct measurements of gravity waves from the troposphere up to the mesosphere. Other remote sensing instruments such as radars are not able to access this height range.

The Esrange lidar (68° N, 21° E) [Blum and Fricke, 2005] is especially suitable to participate in this task since it provides one of the longest available time-series of observations in the Arctic. The lidar was installed at Esrange in winter 1996/97. Hence the retrieved data set covers more than 18 years of measurements. The lidar is located east of the Scandinavian mountain ridge and thus under a strong influence of orographically induced gravity waves, so called mountain waves. We will therefore evaluate the statistical nature of the gravity wave activity measured by the Esrange lidar together with its seasonal behavior. Also we will discuss differences between our measurements and previously published studies.

Since instruments have limitations in spectral and temporal resolution [Gardner and Taylor, 1998] multi-instrumental and multi-stationary measurements are needed in order to resolve gravity waves on all scales. Thereby a comprehensive climatology of gravity waves can be produced [Fritts and Alexander, 2003]. Following this approach, the Gravity Wave Life Cycle campaign was conducted in December 2013 in northern Scandinavia which is also relevant for this thesis. The main goal of the campaign was to combine in-situ measurements, ground-based observations and model studies in
order to study the life cycle of gravity waves as well as their propagation from the
troposphere into the middle atmosphere. We contributed by measuring perturbations
associated with gravity waves from the stratosphere up to the lower mesosphere.

This thesis starts with describing the theoretical background in chapter 2. Chapter 3
discusses the experimental setup of the Esrange lidar and the newly developed method
used for studying gravity wave parameters. After that the results are shown in chapter
4, which are then discussed in detail in chapter 5. Finally, chapter 6 presents a summary
of the findings.
Chapter 2

Theoretical background

2.1 Structure of the atmosphere

The Earth’s atmosphere can be divided into several layers of different physical nature. The transitions between these layers are defined by a change of sign in the temperature gradient (Figure 2.1):

The lowest layer is the so called troposphere. It is characterized by a decreasing temperature with increasing altitude. It reaches up until approximately 10 km. The troposphere is dominated by turbulence and weather phenomena.

The stratosphere (10 to 50 km) is located above the troposphere and shows a temperature increase with altitude. This is due to the absorption of solar ultraviolet radiation by ozone in this altitude range. Because of this positive temperature gradient the stratosphere is also rather stable and stratified (hence the name). In the lower stratosphere polar stratospheric clouds can be observed close to the winter pole [McCormick et al., 1982]. Another prominent feature of the stratosphere is the Junge layer which is a layer of aerosols at an altitude range between 10 and 30 km [Junge et al., 1961].

At 50 km altitude the mesosphere begins which in turn has a negative temperature gradient. The mesopause is located either at 85 km or between 95 – 100 km depending on latitude and time of the year [Xu et al., 2007]. In the summer mesosphere noctilucent clouds can be observed [Stebel et al., 2000] as well as polar mesospheric summer echoes [Rapp and Lübken, 2004]. On top of the mesosphere lies the thermosphere where the temperature is strongly increasing with height.
Theoretical background

But not only the temperature changes within the different atmospheric layers as can be seen from Figure 2.1. Also the pressure $p$ and the density $\rho$ are a function of the height $z$. The link between these variables is given by the hydrostatic equation [e.g. Wallace and Hobbs, 2011]:

$$\frac{\partial p}{\partial z} = -\rho g$$  \hspace{1cm} (2.1)

Where $g$ is the acceleration due to gravity.

By inserting the ideal gas law into equation 2.1 and solving it for $p$ one can easily derive the so called barometric formula:

$$p(z) = p(z_0) \exp\left(-\frac{z - z_0}{H_s}\right) \quad \text{where} \quad H_s = \frac{k_B T}{mg}$$  \hspace{1cm} (2.2)

Here $k_B$ is Boltzmann’s constant, $T$ the temperature and $m$ the mean molecular mass of air. As the density $\rho$ is directly proportional to the pressure $p$, one can easily substitute one for another in equation 2.2. Examining the barometric formula gives the result that pressure and density are both decreasing exponentially with height. Note that equation 2.2 assumes that the atmosphere is isothermal. Although this is

![Figure 2.1: Mean temperature profiles in the northern hemisphere during January and July taken from the MSIS 86 model (Hedin [1991])](image)

generally not true, equation 2.2 still remains a very effective tool to examine basic features of the atmosphere.

2.2 Middle atmospheric circulation

The mean circulation in the middle atmosphere is driven by geographical differences in solar heating as well as forcing due to breaking waves (see section 2.3). The zonal mean zonal wind is generally eastward during winter and westward during summer (Figure 2.2). At an altitude of approximately 90 km this is reversed and the mean zonal wind is directed westward during winter and eastward during summer [Brasseur and Solomon, 2005].

In the winter stratosphere the so called polar vortex or polar night jet is formed. This vortex arises during the polar night when there is no solar heating since there is no absorption of UV light due to ozone in the stratosphere. Therefore the air cools and subsidizes which leads to the formation of a strong westerly jet in the stratosphere. The polar vortex starts developing in fall and reaches its maximum strength in the mid of winter. During late winter and early spring the polar vortex decreases in strength and stratospheric warmings become more likely. These warmings are associated with breaking planetary and gravity waves which trigger an increase

![Figure 2.2: Mean zonal mean wind speed in the northern hemisphere during winter and summer and its effect on vertical wave propagation (see section 2.3 for more details), after Lindzen [1981]](image-url)
in stratospheric temperature and a reversal of the wind [Duck et al., 2001; Shepherd, 2000].

The mean meridional circulation can be distinguished into two branches. One branch is the wave-driven Brewer Dobson circulation which is located at stratospheric altitudes. It originates in the tropics and transports tropical air into high latitude regions in both hemispheres. The other branch is located at altitudes around the mesopause. This is driven by dissipating gravity waves (see section 2.3) and transports air from the winter pole towards the summer pole.

2.3 Gravity waves

As there exists a strong link between the mean middle atmospheric circulation and atmospheric waves we want to examine these waves more closely. We focus thereby on gravity waves. These waves are generated if an air parcel with density $\rho_p$ is adiabatically displaced by a distance $\delta z$ from its position of rest. The buoyancy force then acts as a restoring force on the air parcel which can be described by the following equation:

$$\rho_p \frac{d^2(\delta z)}{dt^2} = -g(\rho_p - \rho_a) \tag{2.3}$$

Where $\rho_a$ is the density of the air surrounding the air parcel.

This equation can be converted into the equation for an harmonic oscillation with frequency $N$:

$$\frac{d^2(\delta z)}{dt^2} = -N^2 \delta z \tag{2.4}$$

With the so called Brunt-Väisälä Frequency $N$:

$$N = \sqrt{\frac{g}{T} \left( \frac{\partial T}{\partial z} + \frac{g}{c_p} \right)} \tag{2.5}$$

Where $T$ is the temperature of the atmosphere, and $c_p$ is the specific heat capacity of air at a constant pressure.

From equation 2.4 one can also see that this oscillation is only possible if $N^2 > 0$ which is also the criterion for a stably stratified background atmosphere. Otherwise
the motion will grow exponentially which leads to an instability. This is the case if the buoyancy force acts in the direction in which the air parcel is displaced.

These oscillations can now propagate not only horizontally but also vertically. In order to capture this feature with an equation one has to look at the fundamental fluid equations and use a linear perturbation theory on them. Assuming a background atmosphere which is in hydrostatic balance and applying the Boussinesq approximation one can derive an equation of motion for gravity waves that can propagate in all three dimensions. A detailed description of this derivation can be found in Fritts and Alexander [2003] or Nappo [2002].

As the fundamental fluid equations follow from the conservation of mass, momentum and energy, the equations describing gravity waves and therefore the waves themselves have to have the same features as well. This leads to an interesting phenomenon: In absence of dissipational processes, the waves kinetic energy has to be conserved. The kinetic energy in turn is proportional to the waves amplitude squared. The amplitude itself is inversely proportional to the density which due to equation 2.2 decreases exponentially with height. As a result the amplitude of a gravity wave has to grow exponentially with height [e.g. Brasseur and Solomon, 2005]. This is one of the reasons why the impact of gravity waves on the atmospheric circulation becomes larger with increasing altitude. But if the amplitude becomes too large, the wave becomes unstable and breaks. Breaking waves add their energy and momentum to the background atmosphere and thereby influence the direction and amplitude of the background winds. Further, breaking induces turbulence which in turn mixes the atmosphere. This becomes especially important in the mesosphere where gravity wave breaking is considered to be the most important source of turbulence [Lindzen, 1981]. This turbulence is for example partly responsible for the existence of polar mesospheric summer echoes which are exceptionally strong radar echoes in the VHF band (30 to 300 MHz) [Rapp and Lübken, 2004].

But gravity waves also influence other parts of the atmospheric system: By inducing temperature anomalies for example, they can enable the formation of clouds such as polar stratospheric clouds [Dörnbrack et al., 2002]. Also they have an impact on atmospheric chemistry as can be seen for example from airglow measurements [e.g. Walterscheid et al., 1999].

Another phenomenon that can be observed is critical level filtering of gravity waves as described by Lindzen [1981]. A gravity wave approaches a critical level if its intrinsic
horizontal phase speed is equal to the horizontal wind speed of the atmosphere. The intrinsic phase speed is the speed with which the phase of the wave moves relative to the background wind speed. If a wave experiences such a critical level it gets absorbed and transfers all its momentum and energy to the mean flow. Looking at the mean wind profile (Figure 2.2) which changes during the year it becomes evident that during winter mostly westward propagating gravity waves can reach the middle atmosphere. During summer this is reversed as the mean zonal wind speed changes sign. Thereby it becomes evident that breaking gravity waves result in a westward directed forcing of the middle atmosphere during wintertime whereas they result in an eastward forcing during summertime [Brasseur and Solomon, 2005]. This seasonal difference in gravity wave forcing drives the inter-hemispheric circulation in the mesosphere [e.g. Holton and Alexander, 2000] which results in the unusually cold polar summer mesopause. Therefore gravity wave filtering processes are crucial to the thermal structure of the polar mesosphere.

As described in the beginning of this section, in order to get a wave like motion an air parcel has to be displaced from its position of rest. There are different sources that can account for this. The main sources are considered to be excitation by topography, convection, shears, geostrophic adjustment or wave-wave interactions. The most well understood source is topography. If the wind blows for example towards a mountain ridge the air is forced to move upward. The buoyancy force will act on the air that was forced upward and an oscillatory movement starts like described before if the atmosphere is stably stratified. The waves generated in this process are also referred as mountain waves. The excitation mechanisms associated with the other gravity wave sources are not so well known. Often several possible mechanisms have been proposed to explain the generation of gravity waves by these events. A detailed description of these mechanisms can be found e.g. in Fritts and Alexander [2003].

One way to quantify the gravity wave activity is to calculate the gravity wave potential energy (GWPED) per volume $E_{pot,V}$ which should be approximately constant with height if there is no dissipation [Rauthe et al., 2008]:

$$E_{pot,V} = \bar{\rho} - \frac{1}{2} \frac{g^2}{N^2} \left( \frac{\hat{\rho}}{\bar{\rho}} \right)^2 \approx \bar{T} - \frac{1}{2} \frac{g^2}{N^2} \left( \frac{\hat{T}}{\bar{T}} \right)^2$$

(2.6)

With the fluctuations of density $\hat{\rho}$ and temperature $\hat{T}$, the mean density $\bar{\rho}$ and temperature $\bar{T}$ and the Brunt-Väisälä frequency $N$ from equation 2.5.
A higher gravity wave activity would induce stronger density (temperature) fluctuations from the background density (temperature). Thus a higher GWPEP per volume can be observed as can be seen from equation 2.6. Note that some studies [e.g. Yamashita et al., 2009] also show GWPEP per mass \( E_{pot,M} = \frac{E_{pot,V}}{\rho} \) which in the absence of dissipation increases exponentially with height. This is due to the fact that the density decreases exponentially with height (compare equation 2.2) and \( E_{pot,M} \) is proportional to \( \frac{1}{\rho} \). Rauthe et al. [2008] for example show a GWPEP per volume of 0.03 J/m^3 at 30 km during winter which is higher compared to 0.015 J/m^3 during summer. Satellite studies also show a larger gravity wave activity at mid and high latitudes during winter than during summer [Alexander et al., 2008].

One difficulty when measuring gravity waves is the problem of comparing measurements of different instrument types to each other. [Gardner and Taylor, 1998] state that each instrument observes only a distinct part of the gravity wave spectrum due to the different measurement principles the instruments are based on. Also different instruments measure at different altitude regions: While lidars are capable to conduct one dimensional measurements from the troposphere up to the mesopause, radars can only measure in the troposphere and in the mesosphere/lower thermosphere and airglow measurements can only be conducted at altitudes around the mesopause. Satellite measurements, on the other hand, do not have such a good temporal resolution as ground based instruments but can provide a unique horizontal coverage. Therefore a combination of all these instrument types is needed for resolving middle atmospheric gravity waves completely.

2.4 Lidar remote sensing

2.4.1 Lidar principles

The acronym LIDAR stands for light detecting and ranging. It is possible to use a ground-based lidar in order to conduct atmospheric measurements with a high temporal and vertical resolution while covering a huge altitude range from the ground up till 100 km. This makes lidars excellent tools to study the middle atmosphere.

The basic setup of a lidar system can be seen in Figure 2.3. It consists of two branches: the transmitting and the receiving branch. The transmitting branch consists mainly of a laser that emits short pulses at a specific
Lasers have several advantages over other light sources which make them ideal for lidar applications: a high power output at a specific wavelength $\lambda$ with narrow spectral bandwidth and very short pulses. Typically Nd:YAG lasers are used for lidar systems which emit light at a primary wavelength of 1064 nm. By frequency doubling and tripling it is possible to generate light at 532 and 365 nm. Thereby multi-wavelength measurements become possible with a single laser. Often the laser beam is expanded before it is emitted into the sky, in order to reduce its divergence.

The receiving branch consists of a telescope that collects all the backscattered light coming from the atmosphere. This light is fed into an optical bench where the time delay between transmitting and receiving the signal and its intensity is measured by photomultipliers. From the runtime the altitude from which the signal originates can be determined whereas the signal intensity yields information about the properties of the atmosphere at this altitude. The signal from the photomultipliers is then processed and saved by a data acquisition system.

Having established the technical basics, one can write the lidar equation which describes the power $P$ received by the lidar as a function of the distance $r$:

$$P(r) = KG(r)\beta(r)T(r)$$

Here $K$ is a factor standing for the performance of the lidar system, $G(r)$ describes the geometry of the experimental setup which in general is proportional to $\frac{1}{r^2}$. $\beta(r)$ is the backscatter coefficient which describes the atmosphere's ability for scattering light back.
to the lidar. $T(r)$ represents the transmission of the atmosphere. It describes how much light gets lost along the light’s path forth and back. A more detailed description can be found in Wandinger [2005].

### 2.4.2 Scattering mechanisms

So far we did not specify how the atmosphere is scattering light back to the receiving branch. In fact there are several different scattering mechanisms that have to be treated individually since each mechanism represents a different physical process. A more detailed description of these processes can be found in Wandinger [2005].

The scattering mechanism most relevant for lidar applications is elastic backscattering due to Rayleigh or Mie scattering. Elastic backscattering refers to the fact that the atmospheric molecules and particles scatter the light back without changing the initial wavelength of the light. By measuring this signal one can retrieve information about the atmospheric density and the presence of aerosols or clouds in the atmosphere. The term Rayleigh scattering describes the situation in which the elastically scattering particles are very small compared to the scattered wavelength. This is generally valid for molecular scattering. The intensity due to Rayleigh scattering is proportional to $\lambda^{-4}$.

The term Mie scattering describes elastic backscattering by particles of all sizes and thus includes Rayleigh scattering. However in atmospheric applications the term Mie scattering is often used only for scattering at particles that are too large for applying the Rayleigh theory. For these particles the backscattered intensity is no longer a function of $\lambda^{-4}$. In fact the intensity can vary a lot with varying wavelength.

Another important scattering mechanism is Raman scattering which is an inelastic process. Raman scattering happens if a molecule absorbs the emitted light but emits it at a different wavelength due to a change in rotational or vibrational quantum number. As the different energy levels are populated depending on Boltzmann’s distribution law, Raman scattering yields information on the atmospheric temperature.

Also resonance scattering can be used for lidar applications. Resonance scattering makes use of the fact that by tuning a laser to a wavelength with an energy that coincides with the energy of an atomic transition in the principal quantum number. This can be used to measure e.g. the Na D$_2$ line between 80 and 110 km where the
atmosphere contains Na atoms.
Additionally one can also make use of the Doppler shift in some scattering signals and thereby retrieve information about winds speeds.
Chapter 3

Experimental setup and analysis method

3.1 The Lidar at Esrange

Since January 1997 a lidar system has been operated at Esrange (68° N, 21° E) close to Kiruna in northern Sweden. Measurements with the Esrange lidar are not conducted continuously but on a campaign basis. Since 2007 only wintertime campaigns were conducted. The lidar is mainly used for stratospheric and mesospheric research since it is able to cover an altitude range from 4 km up to 100 km. It has previously been used for investigating polar stratospheric clouds [e.g. Achtert et al., 2011; Blum et al., 2005], noctilucent clouds [e.g. Stebel et al., 2000], for studies of waves in the middle atmosphere [e.g. Blum et al., 2004], middle atmospheric temperature [Blum and Fricke, 2008], and as a support for rocket and balloon campaigns at Esrange [e.g. Lossow et al., 2009].

This section gives a short description of the technical constituents of the Esrange lidar. A more comprehensive technical description can be found in Blum and Fricke [2005].

The main part of the lidar is a Nd:YAG laser emitting light at a primary wavelength of 1064 nm with a repetition rate of 20 Hz. This primary light is then frequency doubled by a second harmonic generator in order to get a beam with a wavelength of 532 nm. This is the main wavelength used for atmospheric research because not only is the atmosphere almost completely transparent for this wavelength, Rayleigh scattering is also by a factor of 16 more effective at 532 nm than at 1064 nm (compare section 2.3.2). Thus the 532 nm light yields a stronger signal. The 532 nm beam is send through a
telescope and broadened by a factor of 10 in order to reduce the beam divergence. Via a mirror the beam is then send vertically into the atmosphere.

Three Newtonian telescopes collect the light and focus it onto three different focal boxes that separate the received signal according to the wavelength and to the planes of polarization. These signals are then via glass fibre fed into the optical benches designated for the individual signals.

The Esrange lidar has currently four different optical benches: One vibrational Raman channel which can measure number densities of molecular Nitrogen, a rotational Raman channel that has been implemented in 2011 in order to measure atmospheric temperatures below 35 km [Achtert et al., 2013], two channels for measuring the 532 nm light scattered due to Rayleigh and Mie theory, one for the parallel polarized signal parallel and one for the perpendicular polarized signal in respect to the originally emitted light. The latter two signals are cascaded via beamsplitters onto multiple photomultipliers. Each of these photomultipliers has a different sensitivity and records the signal only from a specific altitude range. The rest of the signal for each photomultiplier is blocked by different electrical and mechanical choppers. This procedure is necessary because the signal varies over 7 to 8 orders of magnitude and a single photomultiplier is not able to cover this range. The strong variation of the signal can be explained by the fact that the retrieved Rayleigh signal is proportional to density which is exponentially decreasing with height (compare eq. 2.2) and the $r^2$ dependency of the geometric term in the lidar equation (compare eq. 2.7).

Each photomultiplier is connected to a Joergen counter with a minimum vertical resolution of 1 μs which corresponds to a range gate with a width of 150 m. All these counters are mounted in a computer automated measurement and control (CAMAC) crate which is synchronized with the choppers and the laser. Finally the measured signals are stored by a computer.

### 3.2 Retrieving gravity wave parameters

#### 3.2.1 Density method

For investigating gravity waves with the lidar one has to specify a variable that is affected by gravity waves and that can be deduced from the lidar measurements. One
of these variables is the atmospheric density. The lidar equation (eq. 2.7) can be rearranged to yield the number density of the atmosphere \( N \) as a function of altitude \( z \) and the signal \( S \) measured due to Rayleigh/Mie scattering [Behrendt, 2005]:

\[
N(z) = \frac{CS(z)(z - z_0)^2}{\tau(z_0, z)^2} \quad (3.1)
\]

Here \( z_0 \) is the lidar's altitude above sea level which is equal to 485 m for the Esrange lidar and \( \tau \) is the atmospheric transmission between the lidar and the designated altitude. \( C \) is a constant that includes all parameters that are independent of height. Equation 3.1 assumes that there is no extinction due to clouds and aerosols in the atmosphere. Therefore it is only valid in the aerosol free part of the atmosphere which is approximately above 30 km.

\( \tau \) can be calculated by using the scattering cross section and a density profile from e.g. a model atmosphere. However, we neglect effects of non-uniform transmission for simplicity reasons as this includes a relative error of only 0.1% in the stratosphere which is significantly less than the statistical uncertainties in the measurement process.

The constant \( C \) in equation 3.1 can be calculated by fitting \( N(z) \) at a distinct altitude to a model atmosphere or radiosonde measurements. A radiosonde is a balloon-borne instrument, measuring temperature, wind speeds, humidity, pressure and density as a function of altitude. Since we do not have radiosonde launches available for every day, we use the global thermospheric model based on mass spectrometer and incoherent scatter data (MSIS) 86 [Hedin, 1991]. The signal is normalized to the model atmosphere between 40 and 45 km. The mean value of the normalizing factor is calculated in order to minimize the influences of statistical fluctuations and the perturbations due to gravity waves.

However, the density itself is not of interest for gravity wave measurements. We are first of all interested in the density perturbations. To retrieve the latter we first calculate the mean density by taking the mean of all measured profiles during one night. Afterwards we calculate density profiles for every 15 min. This is done by taking the mean signal over 1 h of measurements and shifting them gradually by 15 min forward in time. From these individual profiles we then subtract the nightly mean profile which is again smoothed by a running mean over 4 km in order to make also stationary gravity wave perturbations visible. We then divide these perturbations \( \tilde{N} \) by the mean density \( \bar{N} \) to get the relative density perturbations \( \frac{\tilde{N}}{\bar{N}} = \frac{\tilde{\rho}}{\bar{\rho}} \). An example for these relative density perturbations can be seen from the measurements during the night at 06-Mar-1998.
Figure 3.1: Relative density perturbations on 06-Mar-1998 between 17:30 and 03:30 UTC between 17:30 and 03:30 UTC (Figure 3.1). Note that the colorbar ranges from 0 to 10% although one would expect also negative density perturbations. This is due to a problem with this method and will be explained later (see section 3.4).

3.2.2 Temperature method

By assuming that the density perturbations can be represented by an adiabatic process one knows that a sudden change in density has to induce a change in temperature as well. Therefore temperature perturbations can also be used for studying gravity waves. One can calculate a temperature profile from the Rayleigh signal by integrating the hydrostatic equation (eq. 2.1) and inserting the ideal gas law. This was first proposed by Hauchecorne and Chanin [1980]. Doing so one ends up with the following equation for the temperature $T$:

$$ T(z) = \frac{M}{k_B} \int_{z}^{\infty} \frac{N(\zeta)}{N(z)} g(\zeta) d\zeta $$

Here $M$ is the molecular mass of air which is assumed to be constant with height and $k_B$ is the Boltzmann’s constant.

Since all measurements reach to a finite altitude, one cannot integrate to infinity. Hence one needs to assume a starting value at a reference altitude $z_m$. Thereby equation 3.2
becomes:

\[
T(z) = \frac{N(z_m)}{N(z)}T(z_m) - \frac{M}{k_B} \int_{z_m}^{z} \frac{N(\zeta)}{N(z)}g(\zeta)d\zeta
\]  

(3.3)

If one then inserts equation 3.1 into equation 3.3 one gets an equation that relates the measured signal \(S\) to the temperature \(T\):

\[
T(z) = \frac{z_m^2 S(z_m)}{z^2 S(z)} T(z_m) - \frac{M}{k_B} \int_{z_m}^{z} \frac{\zeta^2 S(\zeta)}{z^2 S(z)}g(\zeta)d\zeta
\]  

(3.4)

One can see from equation 3.4 that in order to calculate the temperature one has to guess an initial temperature value \(T(z_m)\) at a reference altitude \(z_m\). This can be done by the means of either model data or by a resonance lidar that can retrieve temperature at these altitudes. Also the integration is carried out "top to bottom" meaning that the reference altitude is chosen at the top of the measurements. This is done because the signal \(S(z)\) increases exponentially with decreasing altitude. As a consequence the influence of errors that are made by choosing an incorrect starting value are reduced with decreasing altitude since the first term in equation 3.4 becomes smaller and smaller. Therefore the calculated temperature converges towards the real temperature as altitude decreases.

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**Figure 3.2:** Left panel: Mean temperature profile (dashed line) and individual temperature profiles. Right panel: corresponding temperature perturbations. Both measured at 06-Mar-1998 between 17:30 and 03:30 UTC
Since equation 3.4 is derived by using equation 3.1, it is only valid for aerosol-free conditions as well. This means that this method can only be applied from 30 km upward. However, the Esrange lidar has a rotational Raman channel capable of measuring temperature below this altitude (a detailed description how to calculate the temperature from the rotational Raman signal can be found e.g. in Behrendt [2005] or Achtert et al. [2013]). Nevertheless we do not use the rotational Raman signal for gravity wave analysis since its statistical errors are comparable in size to the calculated temperature fluctuations (see section 3.4).

Similar to the discussion of the density method in section 3.2.1, we are mainly interested in the temperature deviations from the mean temperature of the background atmosphere. The latter is calculated from the nightly mean Rayleigh signal. The initial temperature value $T(z_m)$ at the reference altitude is chosen from the MSIS 86 model at an altitude where the nightly mean counts are by 4 counts higher than the background signal. Note that the counts are obtained by integrating over 5000 shots which corresponds to a time period of 4 min 10 s.

As before we determine time resolving temperature profiles by taking hourly mean profiles shifted by 15 min each. For all individual profiles we use the same initial temperature $T(z_m)$ as for the nightly mean profile. An example for these individual profiles and the nightly mean profile can be seen from the measurements during the night at 06-Mar-1998 between 17:30 and 03:30 UTC (Figure 3.2(a)).

![Figure 3.3: Comparison of different methods for obtaining the mean temperature of the background atmosphere on 06-Mar-1998](image-url)
For the next step the crucial part is how to obtain the mean temperature of the background atmosphere. We therefore tested two different methods: The first was to smooth the nightly mean temperature by applying a running mean over a 4 km altitude region. The second was to apply a smoothing spline fit to the nightly mean temperature profile. The temperature profiles obtained by both methods can be found in Figure 3.3. As can be seen both methods follow the nightly mean temperature rather well. However, the running mean slightly underestimates the temperature around the stratopause. We therefore chose to use the spline fit method further on.

Finally we subtract the background temperature from the individual profiles to determine the temperature deviations from the background temperature. An example for these temperature perturbations can be seen in Figure 3.2(b).

### 3.2.3 Spectral analysis

One of the parameters we want to retrieve from our density and temperature perturbations is the dominant vertical wavelength. This can be achieved by applying spectral analysis methods. We chose to use two different methods: A simple Fourier transformation and a wavelet transformation.

![Wavelet spectrum](a) Wavelet spectrum

![Fourier spectrum](b) Fourier spectrum (blue) and global Wavelet spectrum (red)

**Figure 3.4:** Spectral analysis for the 6th of March 1998. Left panel: Mean Wavelet spectrum, with the amplitudes being color-coded as a function of vertical wavelength and altitude. The black dashed line marks the cone of influence (see text for details). Right panel: Comparison between Fourier and Wavelet transformation
The Fourier transformation is a standard method for spectral analysis. It treats the signal as a sum of sine and cosine functions with different frequencies and amplitudes. As a result one gets the amplitude of the sine or cosine functions as a function of frequency or wavelength. An example of this can be seen in Figure 3.4(b).

The wavelet transformation is more sophisticated [Torrence and Compo, 1998]. It decomposes the signal into the sum of localized wave packets with different frequencies and width. As the Fourier transformation it yields information about which frequencies or wavelengths are dominant in the signal. Additionally it shows in which time or altitude intervals these frequencies are dominant. This information cannot be produced by the Fourier transformation. Figure 3.4(a) shows an example of a wavelet transformation obtained from measurements taken at 06-Mar-1998. One can see the amplitude of different vertical wavelengths as a function of altitude.

Figure 3.4(a) shows also the so called cone of influence (black dashed line). Everything to the right hand side of this line is affected by edge effects due to the fact that there is no measurement of infinite length. It is therefore unclear in how far the information in this area is significant. A more comprehensive description of the wavelet analysis can be found in Torrence and Compo [1998].

To compare the wavelet transformation to the Fourier transformation we calculate the mean spectral power from the wavelet spectrum, which is called the global wavelet spectrum. For comparison both results are shown in Figure 3.4(b). One can see from Figure 3.4(b) that both methods yield similar results. Dominant vertical wavelengths of 6 and 12 km can easily be identified as local maxima in both spectra.

Note that we restrict our spectral analysis to a height range between 30 and 65 km in order to ensure that every analysis has the same limiting factors due to edge effect. By choosing the upper height of 65 km we compromised between a better spectral resolution, being able to apply this method to as many measurements as possible, ensuring sufficient convergence of the temperature towards the true value and a low measurement uncertainty (see section 3.4).

### 3.3 Comparison of the different methods

To validate both methods of retrieving gravity wave parameters an artificial Rayleigh signal with a known wave perturbation was used as a test signal. We know from
equation 3.1 that the raw Rayleigh signal is proportional to the density. Because of this we simulate the raw signal $S_m(z)$ with some random noise $S_n(z)$ from the density $N_m(z)$ of the MSIS 86 model by using the following equation:

$$S_m(z) = \frac{N_m(z) \frac{1}{(z - z_0)^2}}{C} + S_n(z)$$ (3.5)

Where $C$ represents the factor for the system performance. It is set so that it scales the simulated signal to a value that is close to a real signal.

We then add a wavelike signal $S_w(z, t)$ to the raw simulated signal $S_m(z)$:

$$S_w(z, t) = A \sin \left( \frac{2 \pi \lambda_z z}{\lambda_z} + \omega t \right) \frac{1}{(z - z_0)^2} \exp \left( -\frac{z - z_0}{2H_d} \right)$$ (3.6)

Where $A$ is the amplitude of the wave, $\lambda_z$ is the vertical wavelength and $\omega$ the phase speed. The term with the exponential function represents an assumed damping of the wave due to dissipation with a scale height $H_d$. Several different wave compositions were used to test both methods. An example of this can be seen in Figure 3.5.

A first glimpse on Figures 3.5(c) and 3.5(d) reveals no significant difference between the density and the temperature method. We therefore assume that both methods are suited for gravity wave analysis.

Other facts we found by applying several different waves to our model and looking at the spectral decompositions are:

- We cannot observe vertical wavelengths shorter than 2 km. This is due to the smoothing over 1 km altitude. In addition, waves with a wavelength between 2 and 2.5 km are dampened significantly due to the smoothing (not shown here).

- If we apply two different waves in different altitude regions, the wave in the lower altitude region needs a higher amplitude $A$ than the wave in the upper region to reach a similar peak value of spectral power. This can be explained by the fact that the wave amplitude grows exponentially with height. Hence we expect waves in the upper part of the analyzed height range to be more visible in the spectral analysis than waves in the lower part.

- Due to our choice of the height range of the spectral analysis (30 – 65 km) we cannot observe wavelengths larger than $\lambda_{max} = 13$ km, as larger wavelengths are
affected by edge effects of the spectral analysis as shown in Figure 3.4(a) (cone of influence).

So far we assumed that the temperature and the density method both give the same result. We check this hypothesis by looking at the relative density and temperature perturbations obtained from real measurements. We know from equation 2.6 that:

$$\left| \frac{\tilde{N}}{\bar{N}} \right| \approx \left| \frac{\tilde{T}}{\bar{T}} \right|$$

\(3.7\)

![Figure 3.5:](image)

(a) Density perturbations  
(b) Temperature perturbations

(c) Power spectrum of the density method  
(d) Power spectrum of the temperature method

**Figure 3.5:** Upper panel: density (a) and temperature perturbations (b) retrieved from a simulated signal with two waves with vertical wavelengths of \(\lambda_{z,1} = 4\) km and \(\lambda_{z,2} = 7\) km and same amplitude \(A\); Lower panel: corresponding Fourier (blue) and Wavelet power spectrum (red). Note that the amplitude scale in the left panel shows relative units and in the right panel absolute units.
Experimental setup and analysis method

Figure 3.6: Comparison between the relative temperature (left) and density perturbations (right) on 06-Mar-1998 between 17:30 and 03:30

So generally both temperature and density perturbations are expected to show the same behavior. Figure 3.6 shows both perturbations derived from the measurements at 06-Mar-1998. One can see that the relative density perturbations of all profiles show the same amplitude between 40 and 45 km. This happens to be exactly the height where we normalize our signal. This feature always appears in the density profiles at the height range at which we normalize the signal to the reference atmosphere. Also the relative density perturbations above and below the fitting altitude range are strongly depending on how the altitude range is chosen. Since this influences the gravity wave analysis significantly, we decided to use the temperature method further on. The necessity of choosing an initial temperature values for the integration only influences the uppermost temperature values (compare section 3.4). We therefore exclude the highest altitudes from the spectral analysis.

3.4 Measurement uncertainties

We also show the measurement uncertainty as can be seen for a nightly mean temperature profile in Figure 3.7. The measurement uncertainties comprise the effects of the photomultipliers dark count rates, the standard deviation of the measurements and the influence of the signal to noise ratio. The mean temperature profiles shown in Figure 3.7 were obtained on 03-Dec-2013 and integrated over 7 hours. Thus the mean profiles
Figure 3.7: Mean temperature profile and corresponding measurement uncertainty. The profiles were obtained on 03-Dec-2013 and integrated over a time span of 7 hours.

Measurement uncertainty is smaller than the uncertainty of the profiles averaged over 1 hour. The latter’s measurement uncertainties are approximately higher by a factor of $\sqrt{7}$.

Note that Figure 3.7 shows the mean temperature profile obtained by the rotational Raman channel as well. One can see that the measurement uncertainty in the rotational Raman channel goes up to 8 K at an altitude of 30 km. This can be explained by the fact that the count rate of the lidar is proportional to the density of the atmosphere and thus decreases exponentially with height. A lower count rate means higher uncertainty of the measured value hence the increase in uncertainty with altitude. The same is true for the integration technique as well. As the uncertainties for the rotational Raman channel are larger than typical temperature perturbations retrieved at these altitudes, we do not use the rotational Raman signal for gravity wave analysis.

Note also that Figure 3.7 shows only the measurement uncertainty and does not include systematic errors. Systematic errors comprise methodological and instrumental errors. To minimize the instrumental errors the Esrange lidar’s optical bench is recalibrated on a routine basis. The positions of the photomultipliers is adjusted as well as the position of the optical fibers in the focal boxes. Also we ensure that the polarization cubes in the focal boxes are correctly positioned. Additionally the temperature in the detector room is kept stable in order to ensure that the transmission of the interference filters does not change during the measurements.
For the methodological error of the integration technique, we can give an estimate: We know from equation 3.4 that the main methodological error comes from the estimated initial temperature value. Note, that with increasing density this error becomes smaller. The density increases exponentially with decreasing altitude with the atmospheric scale height of \( \approx 7 \) km. By assuming that in the presence of gravity waves the initial temperature could have been wrong by up to \( \pm 20 \) K while starting to integrate at 70 km we can conclude that at an altitude of 65 km this error has decreased to \( \pm 10 \) K and at 49 km it has even decreased down to \( \pm 1 \) K. This starting value problem is the reasons why we discard the uppermost values of the temperature profiles in order to get a good balance between reliable values and spectral resolution (compare section 3.2.3).

Note, that the absolute uncertainties in the temperature profile obtained by the integration technique become higher at 65 km altitude than the uncertainties in the rotational Raman channel at 30 km. Nevertheless we still use the integration technique for wave analysis. This is due to fact that the amplitude of wave-induced temperature perturbations grows exponentially with altitude and typically reaches values of \( 15 – 20 \) K around 65 km altitude (compare Figure 3.2(b)). Thus the temperature perturbations are larger than the measurement uncertainties and are therefore significant.
Chapter 4

Results

4.1 Esrange lidar data

With the previously described method of obtaining gravity wave parameters we analyzed all the data obtained by the Esrange lidar between October and March starting from winter 1996/97 up to the winter 2013/14. This leaves us with 386 days of measurements obtained during a time span of 18 years. However not all of them are suitable for the gravity wave analysis. Since the signal has to reach up to an altitude higher than 65 km, only 213 days of measurements could be used. An overview over these days can be seen in Figure 4.1.

![Figure 4.1](image)

**Figure 4.1:** Overview over all the data used for the gravity wave analysis. Measurement campaigns are marked in blue and suitable measurements for gravity wave analysis with diamonds. Courtesy of P. Achtert
Since the signal strength and thus the observable height range, can vary quite strongly during a measurement due to changes in cloudiness or laser signal strength, we had to ensure that the signal to noise ratio at 65 km was high enough for starting the integration technique at this altitude or above. Thus only those times of the day were used when the mean counts in 40 km height were higher than 450 counts integrated over 5000 shots. Additionally a measurement has to be at least 2 hours long in order to derive gravity wave perturbations from the background profile. This left us in total with \( \approx 1500 \) h of data for the analysis.

Since the background temperature in the middle atmosphere can change dramatically in a matter of hours, we also had to ensure that we always analyzed periods where the atmosphere was in a steady state. Otherwise we would get strong temperature perturbations which are not due to gravity waves but changes in the synoptic situation or due to tides. In these cases we partitioned the measurement period into multiple periods during which the atmosphere was in a steady state and the atmospheric temperature did not vary extensively. This procedure is especially important since Esrange is located at the edge of the polar vortex [see Harvey et al., 2002]. Hence it can happen that at one time of a measurement the observed volume lies within the polar vortex and thus experiences lower temperatures and some hours later the observed volume lies outside the polar vortex and thus a higher background temperature would be measured.

### 4.2 Long-term statistics

As said before we analyzed the data from the last 18 years of measurements during wintertime. The statistical results are shown in this section.

One thing we looked at is the occurrence frequency of different vertical wavelengths \( \lambda_z \) which can be seen in Figure 4.2. This figure was produced by taking the peaks from the global wavelet spectra for every single profile. Thus we only looked at the dominant vertical wavelengths in the spectra. As discussed in section 3.3 we can only detect vertical wavelengths with lengths between 2 and 13 km due to our finite observable height range. Figure 4.2 shows that all the wavelengths observed during the analysis show almost the same occurrence frequency. Only the short wavelengths show a significant lower occurrence frequency since they are dampened due to our method of obtaining the temperature perturbations (compare section 3.3).
Figure 4.2: Occurrence frequency of different vertical wavelengths as obtained from the global wavelet spectrum

Figure 4.3 shows the mean GW PED per volume during different months between 30 and 65 km altitude. The GW PED per volume should be constant with height if there is no dissipation of wave energy (see section 2.3). One can see from Figure 4.3(b) that this is the case during all months above a height of approximately 45 km. Below this height the energy increases by almost one order of magnitude. This dissipation height is generally lower than the monthly mean stratopause height which is denoted by the

Figure 4.3: Mean gravity wave potential energy density per volume during different months. The crosses mark the mean stratopause height during the respective month obtained from the lidar measurements. The numbers in brackets denote how many days of data were analyzed in order to get the mean profiles.
crosses in Figure 4.3. We calculated the monthly mean stratopause height from the lidar measurements by assuming that the stratopause coincides with the height of the temperature maximum in the nightly mean temperature profile between 30 and 65 km. Note that the stratopause can be located higher than at 65 km altitude [e.g. Yamashita et al., 2013], but this cannot be resolved by our method of observation. However, we find no indication for this being the case in our data set.

Figure 4.3 also shows that there is generally a higher GWPED per volume during November and December than during January and February. To see this effect more clearly we also show the mean values of the GWPED per volume between 30 and 40 km and between 40 and 50 km in Figure 4.4. As one can see, the GWPED shows a decrease in both altitude regions during January and February. However the GWPED increases from February to March between 40 and 50 km altitude while it decreases between 30 and 40 km altitude. Figure 4.3(a) shows the same feature, since the GWPED per volume during March is up to an altitude of 37 km higher than the GWPED in January and February. Above 37 km the GWPED during March is generally lower than during the rest of winter.

As described above, we determined the monthly mean stratopause height from our measurements. The results can be seen in Figure 4.5. We find that the monthly mean stratopause is located between 47 km and 51 km. The highest mean stratopause height

![Figure 4.4](image-url)

**Figure 4.4:** Mean gravity wave potential energy density per volume during different months for different altitude regions (note that the figures have a different scale on the y-axis). The numbers of analyzed days are the same as for figure 4.3.
Figure 4.5: Monthly mean stratopause height observed by the Esrange lidar. The numbers of analyzed days are the same as for figure 4.3.

(51 km) is observed in November. It then decreases throughout the winter and reaches its lowest altitude in February (47 km). It then increases back to 49.5 km in March.

Since the monthly mean stratopause height and the monthly mean GWPED per volume between 30 and 40 km show a similar behavior (compare Figure 4.5 and Figure 4.4(a)) we calculated the correlation between the GWPED per volume and the stratopause height. By doing so we find a positive linear correlation coefficient of $r = 0.91$ and a significance level of $p = 0.04$ between the monthly mean stratopause height and the monthly mean GWPED per volume between 30 and 40 km. Calculating the correlation between monthly mean stratopause height and the monthly mean GWPED per volume between 40 and 50 km we get a linear correlation coefficient of $r = 0.50$ and a significance level of $p = 0.39$.

Thus there is a strong significant correlation between the stratopause height and the mean gravity wave activity between 30 and 40 km. However there is no significant correlation between the stratopause height and the mean gravity wave activity between 40 and 50 km.

To determine if there is any trend visible in gravity wave activity during the years we calculated the mean GWPED per volume between 30 and 40 km and 40 and 50 km during January for all years (Figure 4.6). In order to have a representative dataset we limited ourselves to January since this is the month with most measurements (compare Figure 4.1).
One can see from Figure 4.6 that the GWPED varies strongly during different years in both altitude regions but no clear trend is visible from our data set.

### 4.3 Comparison to the ECMWF operational analysis

We also checked the agreement between our measurements and the European Centre for Medium-range Weather Forecast (ECMWF) operational analysis. Therefore we calculated the stratopause height and the stratopause temperature from the ECMWF analysis and compared them to our measurements during winter 2001/02, 2003/04 and 2013/14. We chose these winters since they cover the largest time spans in our data set. We could not compare the winter 1996/97 since the ECMWF analysis from these years does not reach up to the stratopause. Since there are only values for every 6 hours in the ECMWF analysis we also calculated the stratopause temperature and stratopause height for every 6 hours from the lidar measurements in order to have comparable values. An example for the winter 2013/14 is shown in figure 4.7.
Results

Figure 4.7: Comparison of stratopause height and stratopause temperature between the ECMWF operational analysis and the lidar measurements during winter 2013/14. The crosses mark the actual values whereas the lines denote a running mean over the actual values.

One can see from this figure that the lidar measurements generally differ from the ECMWF values. Figure 4.7(a) shows that the stratopause as seen by the lidar is higher in the end of November and beginning of December 2013 than the ECMWF stratopause. After the 7th of December the lidar stratopause is lower than the ECMWF stratopause. In January 2014, on the other hand, they are in a better agreement. However, in the second half of January the lidar stratopause is again higher than the stratopause seen by ECMWF.

In order to quantify this behavior we calculated the mean difference in stratopause height between the lidar and the ECMWF operational analysis (Table 4.1). We found that the difference between lidar stratopause and ECMWF stratopause is on average less than 1 km. During winter 2001/02 the lidar stratopause is on average lower than the ECMWF stratopause, while it is higher during winter 2003/04 and 2013/14. In general

<table>
<thead>
<tr>
<th>Year</th>
<th>$H_{S,lid} - H_{S,ECMWF}$ [km]</th>
<th>$T_{S,lid} - T_{S,ECMWF}$ [K]</th>
</tr>
</thead>
<tbody>
<tr>
<td>2001/02</td>
<td>$-0.9 \pm 5.8$</td>
<td>$16.5 \pm 15.9$</td>
</tr>
<tr>
<td>2003/04</td>
<td>$0.1 \pm 14.6$</td>
<td>$7.1 \pm 9.0$</td>
</tr>
<tr>
<td>2013/14</td>
<td>$0.5 \pm 3.8$</td>
<td>$9.4 \pm 10.2$</td>
</tr>
</tbody>
</table>

Table 4.1: Difference in stratopause height and stratopause temperature between lidar measurements and the ECMWF operational analysis for several years.
the stratopause temperature seen by ECMWF is lower than the temperature measured with the Esrange lidar as can be seen from Table 4.1. The maximum difference in stratopause temperature is seen in winter 2001/02 during which the lidar stratopause is on average 16.5 K warmer than the ECMWF stratopause.

### 4.4 The GW-LCYCLE campaign

Starting from 1st of December 2013 until 14th of December 2013 the GW-LCYCLE campaign took place in northern Scandinavia. GW-LCYCLE stands for Gravity Wave Life Cycle. The goal of this campaign was to study the life cycle of gravity waves and their propagation from the troposphere into the middle atmosphere in northern Scandinavia by combining in-situ measurements, various remote sensing instruments and modeling studies. It combined radar and lidar measurements at Andøya (69° N, 16° E) and Esrange (68° N, 21° E), as well as coordinated radiosonde launches from Andøya, Kiruna Airport (68° N, 20° E), Esrange and Sodankylä (67° N, 27° E). Additionally the German research aircraft Falcon operated during some intensive observation periods (IOPs) in the airspace between Andøya and Sodankylä and thus passed by close to Esrange. Figure 4.8 shows an overview plot of measurements conducted at Esrange from 24-Nov-2013 to 15-Dec-2013. It shows the temperature difference from the nightly mean background temperature measured by the Esrange lidar as well as the times of radiosonde and ozonesonde launches from Esrange. Additionally the IOPs are marked as well. All the times where there are no lidar measurements depicted in Figure 4.8 there were either no measurements possible due to the weather or the signal to noise ratio was not good enough for applying our method of temperature retrieval. Note that some of the lidar measurements had to be subdivided since the background temperature varied strongly during these times. These transitions are marked by the black dashed lines in Figure 4.8.

The times of the IOPs were chosen due to different interesting meteorological situations: During IOP 1 on 3rd of December deep mountain wave propagation was expected due to a strong eastward flow perpendicular to the Scandinavian mountain ridge. IOP 2 was scheduled for the 5th of December in order to investigate gravity wave propagation across the tropopause induced by the jet stream and a low pressure system over the middle of Sweden. During IOP 3 on the 9th of December no significant mountain wave excitation was expected. The focus during this IOP was set on trace gas
Figure 4.8: Overview of measurements from 24-Nov-2013 to 15-Dec-2013; the color-coded panel shows the temperature difference from the nightly mean background temperature measured by the Esrange lidar whereas the black dashed lines mark transitions between steady states of the atmosphere; the red lines mark the time of radiosonde launches from Esrange and the blue line marks an ozonesonde launch; the gray shaded areas mark the flight times of the Falcon aircraft.

measurements which were believed to originate over Asia. On the 13th of December (IOP 4) mountain waves were expected to be excited due to south-eastward flow across the Scandinavian mountain ridge. Another IOP was originally planned for the 11th of December during which strong eastward winds were expected to excite mountain waves at the Scandinavian mountain ridge. This IOP had to be canceled since the strong ground winds made a start of the Falcon aircraft too dangerous. However, radiosondes were launched every three hours from Andøya, Esrange and Sodankylä and ground-based measurements were conducted at Andøya and Esrange.
Table 4.2: Mean GWPED per volume between 30 and 40 km measured at the days of the IOPs during the GW-LCYCLE campaign. The times marked with a, b, c correspond to the different stable regimes during the campaign, separated by the black dashed lines in Figure 4.8.

Table 4.2 shows the mean GWPED per volume between 30 and 40 km. One can see that the highest GWPED was reported at the 11th of December which corresponds to the canceled IOP. The very strong ground winds are believed to yield a strong excitation of mountain waves at the Scandinavian mountain ridge. The GWPED during IOP 4 is also considerably high, indicating a strong excitation of gravity waves as well as favorable propagation conditions. The lowest GWPED was recorded during IOP 2. During this IOP the aircraft measurements took place around 65° N, 20° E where most of the wave activity was expected.

Figure 4.9 shows the occurrence frequency of different vertical wavelengths observed during the IOPs. The wavelengths were calculated from the mean wavelet spectrum between 40 and 50 km since the cone of influence is largest in this altitude region (compare Figure 3.4(a)) and thus the measurement is most accurate.

One can see from Figure 4.9 that during most of the IOPs all vertical wavelengths are observed. However, during IOP 1 we observed mostly waves with a vertical wavelength around 9 km. Whereas there are no longer vertical wavelengths observed during IOP 1. This indicates that this is either a dominant mode of excited waves or the propagation of these waves was more favored than of waves with different vertical wavelengths. During IOP 2 a similar behavior can be seen as there are mainly wave with a vertical wavelength of 7 km observed.

One essential tool during this campaign was the Advanced Research version of the Weather Research and Forecasting (WRF-ARW) model (version 3.4) [Skamarock et al.,
Figure 4.9: Frequency of the dominant vertical wavelengths observed at the days of the IOPs during the GW-LCYCLE campaign. The wavelengths were calculated from the mean wavelet spectrum between 40 and 50 km. The times marked with a,b,c correspond to the different stable regimes during the campaign (black dashed lines in Figure 4.8). Courtesy of P. Achtert

2008] which was used for mesoscale simulations in order to help planning the IOPs. This model was run by Johannes Wagner from the University of Innsbruck with two nested domains with a horizontal resolution of 6 and 2 km (Figure 4.10). The inner domain is thereby computed by one-way nesting. The uppermost level of the WRF simulation is located at 2 hPa, corresponding to about 37 km. To avoid wave reflections at the model top a Rayleigh damping layer was added at the uppermost 5 km. Initial and boundary conditions for the WRF model are supplied by the ECMWF operational analysis on 138 model levels with a temporal resolution of 6 h. A general output of the WRF simulation is possible for every 60 (30) minutes in domain 1 (2). The basic fields like wind, pressure, temperature, water vapour mixing ratio and the Brunt Väisälä frequency are available every 5 minutes for domain 2 (Johannes Wagner, private communication).

The latter outputs were compared to our measurements. To this end, the three dimensional temperature field was interpolated on a one dimensional beam at Esrange like it would have been observed by the Esrange lidar. Additionally we averaged the model output over the same time as the observations by the Esrange lidar. Furthermore we applied the same method of obtaining temperature perturbations to the temperature profiles from the WRF simulations as we used for the lidar measurements. An example of this is shown in Figure 4.11(a). Figure 4.11(a) shows that the WRF simulation
Figure 4.10: The two domains used by the WRF simulations. The red dot marks the location of the Esrange lidar. Courtesy of Johannes Wagner, University of Innsbruck

(below the red dashed line) shows a structure similar to the actual lidar data (above the red dashed line).

In order to check the accuracy of the WRF simulation we plotted the mean temperature profiles obtained by the integration technique and from analyzing the rotational

Figure 4.11: Combination and comparison of lidar data (30 to 65 km) and data obtained by a WRF simulation (0 to 30 km) at 03-Dec-2013. Left panel: Temperature difference from the mean temperature during three different time intervals; above the red dotted line the lidar data is shown and below it the data from the WRF simulation. The black dashed lines show transitions between different stable regimes. Right panel: Mean temperature perturbations deduced from the lidar measurements and the WRF simulation at 03-Dec-2013 between 14:30 and 16:30
Raman channel as well as the mean temperature profile from the WRF simulation between 14:30 to 16:30 (interval a in Figure 4.11(a)) which are shown in Figure 4.11(b). One can see that the temperature profile of the WRF simulation follows the temperature derived from the rotational Raman channel rather well. Only in the upper layers differences can be seen. These differences are within the statistical uncertainty of the rotational Raman channel. Additionally, the WRF simulation and the integration technique show almost the same temperature around 30 km.

Furthermore we applied a wavelet transformation to the combination of temperature perturbations obtained from the lidar data and the WRF simulation which are shown in Figure 4.11(a). The wavelet spectra of these transformations can be seen in Figure 4.12. The upper panel shows the wavelet spectra obtained by analyzing only the lidar measurements whereas the lower panel shows the wavelet spectra obtained by the combination of lidar data and WRF data. One can see that the cone of influence in the figures 4.12(d) to 4.12(f) is considerably larger than for the lidar data only. This is due to the larger height range available for analysis and is purely due to the mathematical nature of the wavelet transformation [Torrence and Compo, 1998].

![Wavelet Spectra](image)

**Figure 4.12:** Mean wavelet spectrum for the three different time intervals shown in Figure 4.11(a) at 03-Dec-2013. The upper panel shows the wavelet spectra derived from the lidar data only whereas the lower panel shows the mean wavelet spectra for the combination of lidar data and WRF simulation.
Comparing the individual figures to one and another one can see that Figure 4.12(a) shows distinct peaks at vertical wavelengths of $6$ and $9$ km. The combination of WRF and lidar data in Figure 4.12(d) shows the same peaks. However, the one at $9$ km reaches down into the lower stratosphere. This indicates that the WRF simulation and the lidar data show the same wave with a vertical wavelength of $9$ km that is propagating from the lower stratosphere up into the upper stratosphere/lower mesosphere. This is the same dominant vertical wavelength that was also recorded in figure 4.9. The wave with the vertical wavelength of $6$ km is not seen by the WRF simulation. This is either due to the fact that this wave could not be resolved by the WRF simulation or that this wave is only propagating at a higher altitude. Figure 4.12(c) and 4.12(f) also show one wave with the same vertical wavelength propagating from the tropopause up to the upper stratosphere/lower mesosphere. Note that Figure 4.12(e) shows a slight increase of the dominant vertical wavelength with altitude from a vertical wavelength of $8$ km at an altitude of $20$ km to a wavelength of $10$ km at an altitude of $50$ km.
Chapter 5

Discussion

In order to evaluate whether there are geographically induced differences in gravity wave forcing, we collected mean values of GWPED obtained by lidars at several different locations from published articles. Table 5.1 shows these values for different altitude intervals. We included lidar studies spanning from the Arctic to the Antarctic. Whiteway and Carswell [1994] analyzed data from a lidar on Eureka, Canada (80° N, 86° W). Blum et al. [2004] compared simultaneous measurements at Andøya, Norway (69° N, 16° E) and Esrange, Sweden (68° N, 21° E) to each other. Thurairajah et al. [2010a] looked at the GWPED at Chatanika, Alaska (65° N, 147° W), while Thurairajah et al. [2010b] looked at Kangerlussuaq, Greenland (67° N, 51° W) and Kühlungsborn, Germany (54° N, 12° E) as well. Data from Kühlungsborn was also analyzed by Rauthe et al. [2008]. Whiteway and Carswell [1995] investigated gravity wave activity over Toronto, Canada during different months. At low latitudes Taori et al. [2012] compared GWPED at Arecibo, Puerto Rico (18° N, 67° W) to GWPED at Gadanki, India (13° N, 79° E). Sivakumar et al. [2006] also analyzed lidar data from Gadanki, India. In the southern hemisphere gravity wave measurements were only recorded at Antarctica. Yamashita et al. [2009] reported values from Rothera (67° S, 68° W) and the South Pole (90° S), while Alexander et al. [2011] reported gravity wave activity measured at Davis (69° S, 78° E).

It becomes evident that most of the studies are conducted in the northern hemisphere. Lidar measurements of GWPED are not available at mid and low latitudes in the southern hemisphere. Also there are only very few long-term statistics of GWPED measured by lidar. The longest records in Table 5.1 are those by Rauthe et al. [2008] (5 years) and Thurairajah et al. [2010b] (4 years).
Looking at the mean values it becomes evident that our values exceed most previous observations. The only values that are higher are those reported by Blum et al. [2004] but these are taken from a case study at Esrange and Andøya. It is likely that single events can show a higher GWPED than the climatological mean. In fact, we recorded single events at Esrange that had a higher GWPED than the one reported by Blum et al. [2004]. Hence we conclude that mean values of single events are subject to strong fluctuations. This can also be seen by looking at the GWPED reported by Taori et al. [2012] that varies over a whole order of magnitude during a few days.

Looking at the distribution of mean GWPED with latitude it becomes evident that the highest values are observed at high latitudes in the northern hemisphere, i.e. our results, those of Blum et al. [2004] and the values associated with no stratospheric warmings recorded by Whiteway and Carswell [1994]. Generally one would also expect the values reported by Thurairajah et al. [2010a] and Thurairajah et al. [2010b] to show a higher GWPED at least those measured at stations north of 60° N. In fact their values of GWPED are the lowest values recorded, together with those of Yamashita et al. [2009] from 90° S. They compare better to those obtained at mid latitudes by Rauthe et al. [2008] and Whiteway and Carswell [1995] which are also lower than those obtained at high latitudes in the northern hemisphere. However, the difference between GWPED measured at Esrange and the one reported by Thurairajah et al. [2010b] may also be due to the fact that Esrange is located downstream of the Scandinavian mountain ridge which could cause a strong local gravity wave forcing. This can also be an explanation for the GWPED reported at Esrange in table 5.1. An explanation for the low values reported by Thurairajah et al. [2010a] and Thurairajah et al. [2010b] is that the density profiles were filtered and smoothed before determining the density perturbations for calculating the GWPED. This procedure reduces the detected density perturbations by a factor of $1.7$ [Thurairajah et al., 2010b] and thus lowers the measured value of GWPED by a factor of $\approx 3$.

Going further to low latitude sites [compare Sivakumar et al., 2006; Taori et al., 2012] a higher GWPED is recorded than at mid latitude sites but still a lower GWPED than the high latitude sites in the northern hemisphere. The results of Yamashita et al. [2009] are hard to compare to all the other sites since they report values from a different altitude region than all the other studies. Compared to our results their values of GWPED are lower. However, those measured at 67° S, 68° W are somewhat comparable to those of Whiteway and Carswell [1994] that are associated with stratospheric
<table>
<thead>
<tr>
<th>Author</th>
<th>Location</th>
<th>period of interest</th>
<th>GWPED per volume [J/m$^3$] 30-40 km</th>
<th>GWPED per mass [J/kg$^3$] 30-40 km</th>
<th>GWPED per mass [J/kg$^3$] 40-50 km</th>
<th>GWPED per mass [J/kg$^3$] 30-45 km</th>
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<tr>
<td>Our results</td>
<td>68° N, 21° E</td>
<td>Nov - Mar</td>
<td>0.203</td>
<td>0.053</td>
<td>30.2</td>
<td>40.2</td>
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<tr>
<td></td>
<td></td>
<td>Nov - Jan</td>
<td>0.198</td>
<td>0.056</td>
<td>30.6</td>
<td>43.7</td>
</tr>
<tr>
<td>Alexander et al. [2011]</td>
<td>69° S, 78° E</td>
<td>May - Sept</td>
<td>0.021</td>
<td></td>
<td></td>
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<td>Blum et al. [2004]</td>
<td>68° N, 21° E</td>
<td>19 - 20/01/03</td>
<td>0.436</td>
<td>0.042</td>
<td></td>
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<tr>
<td></td>
<td>69° N, 16° E</td>
<td>19 - 20/01/03</td>
<td>0.207</td>
<td>0.029</td>
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<tr>
<td>Rauthe et al. [2008]</td>
<td>54° N, 12° E</td>
<td>Nov - Jan</td>
<td>0.026</td>
<td>0.020</td>
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<td>Sivakumar et al. [2006]</td>
<td>13° N, 79° E</td>
<td>Nov - Feb</td>
<td>15.4</td>
<td>31.4</td>
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<tr>
<td>Taori et al. [2012]$^a$</td>
<td>13° N, 79° E</td>
<td>07 - 10/12/09</td>
<td>(2.6) 14.8</td>
<td>(1.6) 29.8</td>
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<td></td>
</tr>
<tr>
<td></td>
<td>18° N, 67° W</td>
<td>07 - 10/12/09</td>
<td>(0.3) 4.8</td>
<td>(0.4) 9.7</td>
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<td></td>
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<tr>
<td>Thurairajah et al. [2010]$^b$</td>
<td>65° N, 147° W</td>
<td>01 - 02/08</td>
<td>1.6</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>67° N, 51° W</td>
<td>01 - 02/08</td>
<td>4.7</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>54° N, 12° E</td>
<td>01 - 02/08</td>
<td>2.6</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Thurairajah et al. [2010]$^b$</td>
<td>65° N, 147° W</td>
<td>Dec - Feb</td>
<td>2.6</td>
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<td></td>
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<td>Whiteway and Carswell [1994]</td>
<td>80° N, 86° W</td>
<td>02 - 03/93$^c$</td>
<td>0.057</td>
<td>0.017</td>
<td>8.7</td>
<td>8.6</td>
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<td></td>
<td></td>
<td>02 - 03/93$^d$</td>
<td>0.112</td>
<td>0.050</td>
<td>14.5</td>
<td>26.0</td>
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<td>Whiteway and Carswell [1995]</td>
<td>44° N, 80° W</td>
<td>01/92</td>
<td>0.025</td>
<td>0.009</td>
<td>13.9</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>03/92</td>
<td></td>
<td></td>
<td>5.24</td>
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<td>Yamashita et al. [2009]</td>
<td>90° S</td>
<td>May - Aug</td>
<td>2.7</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>67° S, 68° W</td>
<td>May - Aug</td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

Table 5.1: Mean GWPED for different altitude regions during winter

- $^a$: The numbers in brackets are the minimum values, whereas the others are the maximum values
- $^b$: GWPED lowered due to data filtering by a factor of $\approx 3$
- $^c$: Cases associated with stratospheric warmings
- $^d$: Cases not associated with stratospheric warmings
warmings. The values measured at the south pole are again significantly lower than those at high latitudes.

In summary one can see that GWPED decreases in the northern hemispheric winter going from high latitudes (larger than $60^\circ$ N) to mid latitudes ($60^\circ - 30^\circ$ N). Low latitude sites (lower than $30^\circ$ N) report a higher GWPED than mid latitude sites but their GWPED is still lower than at high latitude sites. This is consistent with the findings of Alexander et al. [2008] who evaluated the change of zonally averaged temperature perturbations with latitude and height obtained from satellite measurements. Their Figure 6 shows the largest temperature perturbations, and thus, the largest GWPED at high latitudes in the winter hemisphere, a decrease in temperature perturbations in mid-latitudes and a slight increase at low latitudes.

Figure 5.1 shows the occurrence frequency of different values of GWPED for different altitude intervals measured at Esrange between November and March. The occurrence frequencies show bimodal distributions representing two distinct cases of gravity wave activity, indicating two different dominant processes affecting the gravity wave activity at Esrange.

To further quantify this behavior we fitted a curve comprising of two Gaussians to the frequency distributions in Figure 5.1. One can see that the first peak ($xc_1$) is always lower than our mean values shown in Table 5.1, the second peak ($xc_2$) is rather close to our mean values.

The values of $xc_1$ between 40 and 50 km (Figure 5.1, right panel) are close to those reported by Alexander et al. [2011], Rauthe et al. [2008] and the January values reported by Whiteway and Carswell [1995]. Looking at the topography around these three locations it becomes evident that none of them are located close to a strong source of mountain waves. We thus suggest that the first peak in Figure 5.1 can be linked to gravity waves which are excited by ubiquitous sources such as convection, shears, geostrophic adjustment or wave-wave interactions. The second peak on the other hand could be associated with strong mountain wave forcing due to eastward flow over the Scandinavian mountain ridge. Note that the values of $xc_1$ between 30 and 40 km (Figure 5.1, left panel) are close to the values reported by Whiteway and Carswell [1994] for cases associated with stratospheric warmings. But since these cases comprise only the mean of 5 nights of measurement is is arguable in how far they are significant for this comparison.
Figure 5.1: Occurrence frequency of different values of GWPED measured between November and March. Left panel: GWPED between 30 and 40 km; right panel: GWPED between 40 and 50 km. Upper panel: GWPED per volume; lower panel: GWPED per mass. The values xc1 and xc2 mark the center points’ position and uncertainty of the fitted Gaussians (red and green lines). The black line denotes the sum of both curves. Courtesy of P. Achtert

One other feature of Figure 5.1 is that the largest amount of measurements shows a GWPED below our mean value. This is due to the bimodal distribution and the fact that a few measurements with very large wave activity shift the mean towards higher values. Showing the median value would result in a better representation of our measurements. For comparison we show these values in Table 5.2. The median values are almost by a factor of two lower than the mean values. Also they are comparable to the mean values reported by Whiteway and Carswell [1994] for cases not associated with stratospheric warmings.

Figure 4.2 showed the occurrence frequency of different vertical wavelengths, revealing an equally distributed spectrum of vertical wavelengths. This indicates that wave excitation mechanisms around Esrange induce no preferred vertical wavelength. Also
Table 5.2: Median values of GWPED for different altitude regions, derived from measurements with the Esrange lidar.

<table>
<thead>
<tr>
<th>period of interest</th>
<th>GWPED per volume [J/m³]</th>
<th>GWPED per mass [J/kg³]</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>30-40 km</td>
<td>40-50 km</td>
</tr>
<tr>
<td>Nov - Mar</td>
<td>0.091</td>
<td>0.036</td>
</tr>
<tr>
<td>Nov - Jan</td>
<td>0.084</td>
<td>0.038</td>
</tr>
</tbody>
</table>

it indicates that there is no mechanism that selectively prevents waves with a certain vertical wavelength between 2 and 13 km to propagate into the middle atmosphere. Chane-Ming et al. [2000] provided an overview over gravity wave parameters measured at different stations and by different instruments. Our detected range of vertical wavelengths of 2 to 13 km is in good agreement to those of other groups [Chane-Ming et al., 2000, table 2].

Rauthe et al. [2008] present a behavior for GWPED in winter measured with a lidar system at 54° N 12° E between 2002 and 2006. Their Figure 7 is similar to our Figure 4.3(a). However their reported GWPED is approximately one order of magnitude smaller than our values presented in Table 5.1. This can be explained by geographically induced differences.

As our measured GWPED shows the largest change with altitude below the stratopause during all months (Figure 4.3), we conclude that most of the gravity waves’ energy dissipates already below the stratopause either due to critical level filtering or due to the fact that the amplitudes grow to large, and thus, the wave becomes unstable and breaks.

This explains the general form of the GWPED curve throughout the whole winter but it does not explain the lower GWPED during January and February compared to the rest of the winter (compare Figure 4.3 and 4.4). One event that happens generally in January and February is stratospheric warming during which the wind direction in the stratosphere changes from eastward to westward (Figure 2.2). This suggests a different filtering of gravity waves, and hence, a change in GWPED. Thurairajah et al. [2010a] for example measured a lower GWPED during and after stratospheric warming events. Whiteway and Carswell [1994] also reported a lower GWPED in the presence of stratospheric warmings. We thus conclude that this process might be responsible for the lower GWPED during January and February compared to the rest of the winter.
The transition of the mean stratospheric circulation from eastward to westward flow during March decreased the GWPED at high altitudes. This can be seen in an altitude region above 37 km (Figures 4.3 and 4.4(b)), but not below this altitude. A possible explanation for this behavior is that stratospheric warmings influence the stratospheric circulation on shorter timescales than the gradual shift from winter to summer circulation. Like stratospheric warmings the shift from winter to summer circulation propagates downward from the mesosphere into the stratosphere. As this generally happens on larger time scales than stratospheric warmings it is possible to see this effect in our monthly mean profiles of GWPED as well. A different filtering of gravity waves due to the change in background wind, and hence, a change in GWPED would first be visible in a higher altitude region.

Another difference between the altitude regions from 30 to 40 km and 40 to 50 km becomes evident when examining the correlation between the monthly mean stratopause height and the monthly mean GWPED per volume. Most of the GWPED dissipates already below 40 km altitude (Figure 4.4). Dissipating waves add their momentum and energy to the mean flow and thereby influence the mean circulation [Fritts and Alexander, 2003]. One of the responses of the atmosphere to dissipating gravity waves is the deceleration of the polar vortex [Duck et al., 2001]. This in turn has an effect on the stratopause height which can be seen by the strong positive correlation between monthly mean GWPED between 30 and 40 km and monthly mean stratopause (see section 4.2), i.e. a higher gravity wave activity coincides with a higher stratopause.

A comparison of the lidar-derived stratopause height and temperature to the ECMWF operational analysis reveals significant differences. Manney et al. [2008] compared stratopause height and temperature during winter obtained from satellites to ECMWF data. They state that ECMWF shows a higher and warmer stratopause than the observations. This is contrary to our results indicating a lower and colder stratopause in ECMWF data compared to lidar measurements from winter 2002/03 and 2013/14. During winter 2001/02 we find a higher but still colder stratopause in ECMWF data compared to our lidar measurements. Manney et al. [2008] also state that the differences between ECMWF and the satellite data in the higher altitude regions is partly due to the gravity wave parametrization scheme used by the model We found a strong correlation between gravity wave activity and stratopause height. Therefore it becomes very likely that the differences between the lidar measurements and the ECMWF data are due to the insufficient representation of gravity waves in the ECMWF operational analysis.
From Figure 4.6 we deduced no trend of GWPED in our data. Blum and Fricke [2008] on the other hand deduce a trend in gravity wave activity from measurements of GWPED with the Esrange lidar. We argue that this might be an artifact due to their comparable short time range of analyzed data. If only analyzing the years between 1997 and 2005 one could also deduce a negative trend in gravity wave activity from our data set. Looking at larger time spans, this trend vanishes. Thus we conclude that longer time series are needed for resolving trends in GWPED.

We showed in Figure 4.11 and 4.12 that the WRF model reproduces mean temperatures similar to the lidar temperatures. Differences occur only at altitudes above 27 km. This might be either due to large uncertainties in the rotational Raman channel or due to the fact that the WRF model overestimates the local temperature minimum at 27 km altitude (Figure 4.11(b)).

We furthermore deduce from Figure 4.12 that the combination of WRF and lidar data shows gravity waves with similar vertical wavelengths propagating from the lower-most stratosphere up to the upper stratosphere/lower mesosphere. Interestingly the increase in vertical wavelength with altitude observed in Figure 4.12(e) can be explained by the wind shift theory of Eckermann [1995]. It states that if the difference between the horizontal phase speed of the wave and the horizontal background wind speed increases with altitude that the wave is shifted towards a larger vertical wavelength.
Chapter 6

Conclusion

In this Master thesis a method for obtaining gravity wave parameters from lidar measurements was developed. Two methods were tested by applying them to real and artificial data. The temperature method was found to be most suitable as the methodological errors from this method are well characterized and small compared to the density method. The temperature method was applied to the measurements obtained by the Esrange lidar (68° N, 21° E) during wintertime from winter 1996/97 until 2013/14.

We found that the vertical wavelengths detected by the Esrange lidar in wintertime are equally distributed between 2 and 13 km. The upper and lower bound are set by the analysis method. Furthermore we analyzed the monthly mean GWPED. We found that most of the gravity wave energy dissipates well below the stratopause, whereas higher altitude regions show less dissipation of GWPED. The monthly mean GWPED in the altitude region between 30 and 40 km is strongly correlated to the mean stratopause height. Hence we concluded that a strong coupling exists between dissipating gravity waves and the stratopause.

We also compared the stratopause height and temperature determined from the lidar measurements to ECMWF operational forecasts. With the lidar we measured a stratopause that was higher and warmer than the ECMWF stratopause during winter 2013/14. This stands in contrast to the findings of Manney et al. [2008] who deduced a colder and lower stratopause from satellite measurements compared to ECMWF values.

Comparing the mean GWPED measured by the Esrange lidar to other published results we recognized that the highest GWPED was recorded at Esrange. We argued that this is a combination of two different effects: First, the Scandinavian mountain
ridge to the west of the measurement site excites gravity waves and causes a higher GWPED. Second, the GWPED is generally larger at high latitudes which can not only be seen from the lidar measurements (Table 5.1) but was also previously inferred from satellite measurements by Alexander et al. [2008]. Currently only few long-term records of GWPED as measured by lidar systems are available at the moment. In addition, published values might be biased by the way gravity wave perturbations were retrieved. A standardized method of retrieving gravity wave parameters should be used in order to improve the comparability of different studies.

Our measurements show a bimodal distribution in GWPED occurrence frequency. This indicates that two processes affect the gravity wave activity at Esrange. Comparing the position of the peaks in Figure 5.1 to other published studies we speculate that the first peak is due to ubiquitous wave sources whereas the second peak is associated with strong mountain wave forcing. This theory can be tested by looking at the distribution of occurrence frequency of GWPED for lidar systems that are not subject to mountain waves. If they show only a single mode in GWPED occurrence frequency it would support our theory.

Participating in the GW-LCYCLE campaign gave us the opportunity to combine our measurements with mesoscale simulations. We noted a good agreement between the mean temperature measured by the lidar and the one deduced from the WRF simulation. Wavelet transformations showed that gravity waves with the same vertical wavelengths were observed in the lidar data and the WRF simulations. Thus, we argue that the WRF simulation and the Esrange lidar show the same gravity waves. This makes WRF simulations together with our lidar observations a valuable tool to study the propagation of gravity waves from the troposphere up to the mesosphere.

Further studies will be conducted with the data set obtained during the campaign, focusing on excitation mechanisms of gravity waves as well as their propagation into the middle atmosphere and the mechanisms and effects linked to gravity wave dissipation. These issues need to be resolved in order to develop an accurate parametrization of gravity waves in large scale models such as climate models. The need for such an improvement becomes particularly evident when looking at the difference in stratopause height and temperature between the ECMWF operational forecast and our lidar measurements.
Acknowledgements

First of all, I want to thank my supervisor Peggy Achtert: For giving me the opportunity to investigate things I’d find interesting but still pushing me gently towards the important tasks, for letting me work with the lidar by myself and for answering all my stupid questions.

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Finally, my thanks go to my parents, for supporting me during all my studies and for teaching me how to turn a screw.
Bibliography


