Final Report

Technical Assistance for the Deployment of the GLORIA instrument during the Gravity Wave Experiment (GWEX)

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1 Introduction

In the recent years new evidence for the importance of the middle atmosphere for weather prediction and climate projection was presented. For instance, in the review by Kidston et al. (2015) the dynamical coupling of the stratosphere to the troposphere is described: stronger wave drag in the polar vortices increases the meridional mean circulation and shifts, in turn, the position of the jets both in the stratosphere and in the troposphere. Gravity waves are not explicitly described in the paper, but they are among the largest uncertainty factors for accelerations exerted onto the stratospheric winds (e.g. Alexander and Orland, 2010; Geller et al., 2013). A second example is the importance of the Quasi Biennial Oscillation of the stratospheric tropical wind (QBO) for seasonal weather forecasts. It is known, that the surface mean winter temperature in Europe varies 1 to 2 K depending on the phase of the QBO (Marshall and Scaife, 2009). As the QBO itself is a process providing sufficient memory in the atmosphere to enhance seasonal prediction, a good representation of the QBO in weather and climate models is a promising step towards seasonal forecasts (Scaife et al., 2014). The QBO itself is driven to more than 50% by gravity waves (Dunkerton, 1997; Ern et al., 2014). Most recently, Garcia et al. (2017) have shown that missing GW drag at 60° S is responsible for the delayed vortex break down and overestimated ozone loss in the southern polar winter. By an ad-hoc adjustment of their GW launch distribution they can mitigate this problem. However, they only adjust mountain waves on the southern hemisphere leaving the northern hemisphere unchanged. Thus, the mitigation is at the prize of physical inconsistency. Physical inconsistent changes in parametrizations are particularly problematic for climate projections. This is only one example, where modellers use their parametrizations to heal almost any deficiency of the model. They can and have to do that, because the existing satellite climatologies are not sufficient to constrain all free parameters (Fritts and Alexander, 2003). Shepherd (2014) conclude that the largest problems weather prediction and climate projection models are facing are unresolved and, thus, parametrized processes, namely convection in the troposphere and gravity waves in the stratosphere. One of the fundamental problems attacking this issue is the lack of sufficiently accurate global observations, as current day satellites were not designed for these small scales (Geller et al., 2013).

There are a few essential issues for the investigation of GWs, which have to be faced for any observation technique. The first one is that other than temperature or a trace gas mixing ratio GW momentum flux is not a primary observable. In a philosophical sense, GWs are a - very well established and well probed - theoretical concept. In this
concept one has to separate between some atmospheric background and the small amplitude perturbations introduced by the GW. This is the basis for formulating the linear wave theory, but may introduce errors when applied to observations.

The second issue is that no observation technique can resolve all scales of GWs. Even if a sensor (e.g. in situ data) are affected by all scales, also the separation needs to be performed which usually introduces scale limits. For remote sensing instrument the radiative transfer usually smoothes GWs along the line of sight of the instrument. In order to allow quantitative assessments of comparing different instruments or instruments with models, a method is required to quantify this (Alexander, 1998; Preusse et al., 2002; Ern et al., 2006; Ern and Preusse, 2012; Trinh et al., 2016). The observational filter is a measure for the sensitivity of an instrument towards measuring GWs of given horizontal or vertical wavelengths. In general, an observational filter tells by how much the amplitude or the momentum flux is underestimated by the measurement technique.

The best-studied approach for future global observations of GW momentum flux distributions is a space-borne infrared limb imager (IRLI) (Preusse et al., 2009, 2012). This is based on the heritage of the past and current infrared limb scanning missions CRISTA, HIRDLS and SABER which are the corner stone of observational studies on global GW distributions. The IRLI overcomes the major problems associated with today’s space-borne observations, as is listed in table 1.1. It should be noted that also for an IRLI there is still the general observational filter defined by the visibility of GWs to optically-thin limb sounding (applicable also e.g. to GPS). In contrast to current observations, however, GW parameters from an IRLI will be more accurate by factors and some parameters can be determined from an IRLI for the first time. Furthermore, it is obvious (and will be confirmed by this report) that only the regular sampling and tomographic retrieval of an IRLI can elevate the accuracy of limb sounding to the levels required.

Advances in detector technology make an IRLI feasible. The first, airborne demonstrator of this technique is the Gimballed Limb Observer for Radiance Imaging of the

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Table 1.1: Overview of the major error sources of current global observations and the mitigation through the observation geometry of an IRLI.
Introduction

Atmosphere (GLORIA). GLORIA has been developed by the research centers JÜLICH and KIT in a joint effort (cf. Riese et al. (2014); Friedl-Vallon et al. (2014) and companion papers in the GLORIA AMT special issue). For deployment of an IRLI in space, the PREMIER mission was proposed for Earth Explorer 7 (ESA, 2012). In continued effort to demonstrate the technical feasibility and strengthen the scientific case for PREMIER, ESA initiated the Gravity Wave Experiment (GWEX). In this study, the scientific needs for an IRLI for GW research and the scientific advances, which could be gained by such an instrument, are further investigated. The observations of GWEX were performed during the POLSTRACC / GWEX / GW-LCycle / SALSA, in short PGS campaign. Flights targeting GWs took place on 25 January 2016 and 28 January 2016. Flight patterns were designed to allow tomographic retrieval of GWs.

We begin this report with an introduction into the basic theory of GWs required to understand the analyses performed (Chapter 2). This is followed by Chapter 3 which describes the airplane instrumentation, NWP data and tools used in this study. In particular, these are the GLORIA instrument (Section 3.1) and the tomographic retrieval software (Section 3.2), a description of the satellite-based nadir observations (Section 3.3), the ECMWF integrated forecast systems (ECWMF-IFS) from which we used forecast data for campaign planning and analysis data for interpretation of our results (Section 3.4), the tools applied for separation of GW perturbations and background (Section 3.5), the S3D wave fitting tool (Section 3.6), the ray-tracing model GROGRAT used to identify sources and study further propagation of the observed waves (Section 3.7), the flight planning tool MSS used to optimize the measurement flights for our aims (Section 3.8), and ICON mesoscale model simulation to study the various influences of horizontal tracer transport and GW displacement on tracer distributions (Section 3.9).

We have justified the execution of the campaign by a study on the occurrence likelihood of suitable events (Chapter 4) which also served to identify potential target regions. This Chapter also compares the situation in the GWEX winter 2016 with our expectations from prior winters. In order to employ GLORIA in an optimized way and also to quantitatively understand our results, we performed simulation studies described in Chapter 5 for the observation of GWs with different flight patterns allowing full angle tomography (FAT; hexagon) and limited angle tomography (LAT; linear flight). As it is possible to reconstruct 3-D volumes with GLORIA and, thus, derive a 3-D wave vector, we extend the concept of the observational filter. Besides the usual observational filter regarding the wave amplitude, we introduce an observational filter for the wave orientation. This observational filter for the wave orientation with respect to the flight direction is important for the planning of research flights: The wave orientation can often be predicted and the flight path adjusted respectively.

The campaign itself is described in Chapter 6 with a full payload description in Section 6.1 and a discussion of the flight planning in Section 6.2. Two research flights were performed, one Flight on 25 January 2016 over Iceland described in Chapter 7 and a
second flight on 28 January 2016 over Scandinavia in Chapter 8. For both flights the GLORIA observations are described, the observed GWs characterized and the sources investigated. By forward ray tracing we study the further propagation of the waves in the stratosphere. This knowledge is then used to discuss state-of-the-art satellite observations by the Atmospheric Infrared Sounder (AIRS).

GLORIA observes the full 3D wave vector and the temperature amplitude. Based on these the expected wind perturbations can be calculated via the dispersion relation, which in the end leads to the momentum flux. The HALO and Falcon airplane instrumentation observe in situ winds along the flight path. Chapter 9 investigates whether we can bring these data sets together to actually test the dispersion relations, testing the concept first on ECMWF analysis data (Section 9.2) and applying it afterwards on in situ aircraft data (Section 9.3) and dropsonde data (Section 9.4). In this context we also study the impact of GWs on structures observed in ozone and water vapor distributions (Section 9.5).

In the conclusions (Chapter 10) we summarize our findings (Section 10.1) and discuss how such kind of observations could be further optimized (Section 10.2). Finally, in Section 10.3 we discuss the applicability of the study results to a space-borne IRLI and the value of the GWEX project to demonstrate the unique opportunity such an instrument would provide for GW research.
2 Some basics on the theory of gravity waves

A basic understanding of GWs is required to understand the results of this study. In this chapter we largely base on the description of GW physics as given in the final report of the ESA study “Observation of gravity waves from space” performed in support of the PREMIER mission proposal (Preusse et al., 2012). This includes the dispersion relation of GWs, how GWs propagate and interact with the background winds and how they can be observed. In addition to the previous report (Preusse et al., 2012), we here added some equations on displacements and expected GW-induced tracer disturbances. A comprehensive discussion of gravity waves, however, is far beyond the scope of the current report, and for a more comprehensive discussion we refer the reader for instance to Andrews et al. (1987), Gill (1982) or Fritts and Alexander (2003).

2.1 Dispersion relation

Gravity waves perturb atmospheric temperature, density and winds. Their restoring forces are gravity and buoyancy\(^1\). In order to understand the physical principle behind these waves, we start with a thought experiment of an isolated air parcel (cf. Fig. 2.1). If we displace an air parcel upward, it will expand due to decreasing ambient pressure and thus cool. The cooling rate follows the adiabatic temperature gradient \(\Gamma\) which is approximately -10 K/km in the stratosphere. If the temperature gradient of the background atmosphere is larger than \(\Gamma\) (thus positive or less negative), the air in the air parcel will become denser than its surrounding and gravity will force it downward. Vice versa an air parcel displaced downward will be warmer and forced upward. The oscillation frequency of such an air parcel would be

\[
N^2 = \frac{g}{T} \left( \frac{\partial T}{\partial z} - \Gamma \right)
\]

(2.1)

and is called buoyancy frequency or Brunt-Vaissala frequency.

\(^1\)Even though both names, gravity wave and buoyancy wave, were used in the literature, the name gravity wave is established now.
Figure 2.1: Depiction of a thought experiment: Displacing an isolated air parcel upward, the parcel extends, the pressure decreases and the air cools down at a rate $\Gamma = -10 \, \text{K km}^{-1}$. If the background gradient of the ambient air is larger than $\Gamma$, the air becomes denser and is pulled down by gravity. Analogously a downward displaced air parcel becomes lighter and tends to raise. The frequency of the air parcel oscillating around its equilibrium position is called buoyancy frequency $N$.

Many of these oscillating air parcels form a GW. Figure 2.2 gives a sketch of the wind and temperature structure in such a GW. Consider a wave with phases propagating towards the left (and thus downward). We now start, for instance, with a warm phase and, thus, with air which is lighter than its surrounding. Buoyancy will force this air upward and the following phase will upward wind. After the upward movement, the air will be colder and denser than the surrounding. This leads to a downward forcing by gravity and a downward wind. After the downward movement, the air is warmer than the surrounding and lighter, which means buoyancy pushes upward again.

In a GW the wind blows along the phase lines, i.e. each single oscillator moves in the direction of the phase lines. Thus, we can calculate the frequency of a single oscillator and hence the frequency of the wave $\omega$ from the triangle of forces defined by the horizontal wavenumber $k_h = 2\pi/\lambda_h$ ($k_h^2 = k^2 + l^2$ with $k$ and $l$ wavenumbers in zonal and meridional direction) and the vertical wavenumber $m = 2\pi/\lambda_v$. Including the
Some basics on the theory of gravity waves

Coriolis parameter $f$ leads to the dispersion relation

$$\hat{\omega}^2 = \frac{(k^2 + l^2)N^2 + f^2 \left(m^2 + \frac{1}{4\pi^2}\right)}{k^2 + l^2 + m^2 + \frac{1}{4\pi^2}}. \quad (2.2)$$

Using the fact that in the stratosphere $N$ is much larger than $f$, Equation (2.2) can be simplified to

$$\hat{\omega}^2 = \frac{(k^2 + l^2)N^2}{k^2 + l^2 + m^2 + \frac{1}{4\pi^2}} + f^2. \quad (2.3)$$

The wave frequency is limited by $f < \hat{\omega} < N$. The Coriolis parameter $f$ is determined from the Earth rotation frequency $\Omega$ and the latitude $\phi$:

$$f = 2 \ast \Omega \sin(\phi). \quad (2.4)$$

In the mid-frequency range $f \ll \hat{\omega} \ll N$ and for GWs with $\sqrt{k^2 + l^2} \ll m$, Equation (2.3) can be further simplified to

$$\hat{\omega}^2 = \frac{(k^2 + l^2)N^2}{m^2 + \frac{1}{4\pi^2}}. \quad (2.5)$$

Figure 2.2: Depiction of a GW with phases propagating to the left. A warm front precedes upward winds, which precedes a cold front, which precede downward winds, which precedes a warm front. The winds blow along the phase line and a single air parcel (oscillator) in the wave also oscillates along the phase lines. This defines the triangle of forces which directly leads to the dispersion relation.
2.2 Polarization relations and vertical and horizontal tracer displacements

The polarization relation connecting the amplitudes of vertical displacement $\hat{\zeta}$ and temperature $\hat{T}$ is

$$\hat{\zeta} = -\frac{g}{TN^2} \hat{T}$$  \hspace{1cm} (2.6)

where $g$ is Earth’s gravity acceleration and $T$ the background temperature (Eckermann et al., 1998). The wind is the temporal derivative of the displacement. Thus, the vertical wind amplitude $\hat{w}$ can be expressed as follows:

$$\hat{w} = \hat{\omega} \hat{\zeta} = -\hat{\omega} \frac{g}{TN^2} \hat{T}.$$  \hspace{1cm} (2.7)

As the relation of vertical and horizontal wind is given by the ratio of the respective wavelengths, the horizontal wind amplitude $\hat{u}_h$ can be written as

$$\hat{u}_h = \hat{\omega} \hat{\zeta}_h = \frac{m}{k} \hat{w} = \frac{m}{k} \hat{\omega} \hat{\zeta} = \frac{g}{TN} \hat{T},$$

with $\hat{\zeta}_h$ being the horizontal displacement amplitude. To derive this equation we used Equation 2.5 and assumed $\frac{1}{4\pi^2} \ll m$, which is equivalent to making the Boussinesq approximation. For typical stratospheric values one finds that the wind amplitude is about twice the temperature amplitude.

$$\hat{u}_h = \frac{g}{TN} \hat{T} \approx \frac{10}{250 \times 0.02} \hat{T} \left[ \frac{m}{\pi^2 K_s} \right] = 2 \hat{T} \left[ \frac{m}{s} \right]$$ \hspace{1cm} (2.9)

The Earth rotation induces an additional rotational component of the wind, which gets important for waves with intrinsic frequencies $\hat{\omega}$ close to the Coriolis parameter $f$. In this case, the horizontal wind gets divided into components parallel and perpendicular to the wave vector:

$$\hat{u}_\parallel = \frac{\hat{u}_h}{\sqrt{1 - \frac{\omega^2}{f^2}}},$$ \hspace{1cm} (2.10)

$$\hat{u}_\perp = \frac{\hat{u}_h}{\sqrt{\frac{\omega^2}{f^2} - 1}}.$$ \hspace{1cm} (2.11)

The wind component perpendicular to the wave vector is 90° out of phase with the parallel wind component. The ratio of the two wind components is given by the ratio
of intrinsic frequency $\hat{\omega}$ and Coriolis parameter $f$:

$$\frac{\hat{u}_\perp}{\hat{u}_\parallel} = \frac{f}{\hat{\omega}}. \tag{2.12}$$

Thus, taking a vertical wind profile one finds an ellipsis oriented with its main axis parallel to the wind vector. The orientation of the rotation is dependent on the hemisphere and on the fact whether the GW propagates upward or downward. For an upward propagating GW on the northern hemisphere and considering a profile from low to high altitude the rotation is clockwise as shown in Fig. [2.3].

Vertical profiles of the winds can be observed, for instance, by in situ sounding with a dropsonde or radiosonde. For the evaluation of such data fitting hodographs to the observed winds is a frequently-used evaluation technique.

Displacements can be observed by the displacement of tracers and, in particular, if there is a suitable gradient in the tracer. Trace species such as ozone and nitric acid ($HNO_3$) are abundant in the stratosphere and exist only in low concentration in the troposphere, others such as CFCs vice versa. Such species exhibit a strong vertical gradient in the tropopause region and, thus, are suitable for the GW analysis.

A GW leads also to horizontal displacements. In the horizontal, gradients may be found either at mixing-barriers such as the edge of the polar vortex or for filaments which are generated through Rossby-wave breaking and isentropic mixing. In order to assess whether horizontal displacements are relevant we are therefore interested in the displacement of a filament or mixing barrier.

### 2.3 Doppler shift, refraction and propagation direction

The frequency $\hat{\omega}$ is the intrinsic frequency as seen by an observer moving with the background wind. It is related to the ground based frequency $\omega_{gb}$ by

$$\hat{\omega} = \omega_{gb} - k_h U \tag{2.13}$$

where $k_h$ is the horizontal wave vector and $U$ is the horizontal background wind. Using a simplified dispersion relation

$$\hat{\omega}^2 = \frac{k_h^2 N^2}{m^2} \tag{2.14}$$

we obtain a simple expression for the vertical wavelength in dependence of the intrinsic phase speed $\hat{c} = \hat{\omega}/k$

$$\lambda_z = \frac{2\pi}{c - \frac{U_\parallel}{N}} \tag{2.15}$$
Figure 2.3: Hodograph and vertical profiles of an ideal wave. The vertical temperature structure is a sinusoid for given vertical wavelength and temperature amplitude, the other atmospheric variables calculated according to the GW polarization relations. The lower panel shows vertical profiles of all atmospheric variables.
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Figure 2.4: Illustration of critical level filtering. At altitudes where the background wind (blue) matches the phase speed of the upward propagating GWs (green), the vertical wavelength tends to zero and the wave breaks down. At this altitude the waves exert drag in the direction of their propagation (red).

with $U_\parallel$ being the background wind velocity in the direction of the wave vector. When $c$ approaches $U_\parallel$ the vertical wavelength vanishes and the wave cannot further exist / propagate upward. This is called a critical level. The individual waves in a spectrum of upward propagating GWs will therefore reach that altitude at which the phase speed equals the background wind. This critical level filtering is illustrated in Fig. [2.4]

Comparing the phase velocity ($\hat{\omega}/k, \hat{\omega}/l, \hat{\omega}/m$) and the group velocity ($\partial \hat{\omega}/\partial k$, $\partial \hat{\omega}/\partial l$, $\partial \hat{\omega}/\partial m$), we find that a wave packet propagating upward and thus transporting energy and momentum upward has downward phase propagation and that the group velocity points along the phase lines. The group velocity allows us to calculate the path of a wave as it propagates upward in the background atmosphere. Such ray tracing has to take into account that vertical and horizontal wavelength of a wave
are modified by interaction with the background wind as equation (2.15) demonstrates. In particular, vertical wind shear modifies the vertical wavelength, horizontal wind shear the horizontal wavelength and wave direction, and change in time the wave frequency.

### 2.4 Wave drag and pseudo momentum flux

The interaction of GWs with background winds can be quantified by considering the basic equations of atmospheric motions as they are given, for instance, by Andrews et al. (1987), equations (3.2.1) or Fritts and Alexander (2003), equations 1 to 5. The equation for the zonal wind change is

\[
\frac{du}{dt} - f\nu + \frac{1}{\rho} \frac{\partial p}{\partial x} = X
\]

where \( X \) is an unspecified force term and vanishes in a conservative system. These equations can be linearized in a background flow \( \bar{u} \) and and perturbation terms \( u' \) with \( u = \bar{u} + u' \) (all other variables in analogy). Thus we could bring all terms from the background to one side and all terms from the perturbations to the other side, If then all accelerations due to the perturbations are subsumed in \( \bar{X} \) the equation is valid for the background flow, too

\[
\frac{d\bar{u}}{dt} - f\bar{\nu} + \frac{1}{\bar{\rho}} \frac{\partial \bar{p}}{\partial x} = \bar{X}
\]

This means \( \bar{X} \) is now the force term on the background flow and not resolved in this representation, but it is generated by the perturbations (e.g. GWs) to this background flow and can be calculated solving the equation for the perturbations. Using dispersion and polarization relations of GWs it can be shown that in good approximation (Fritts and Alexander, 2003)

\[
\bar{X} = -\frac{1}{\bar{\rho}} \frac{\partial}{\partial z} F_{p,x}
\]

where \( F_{p,x} \) is defined by

\[
F_{p,x} = \bar{\rho} (1 - \frac{f^2}{\hat{\omega}^2}) \bar{u}'w'
\]

and is called the vertical flux of the horizontal pseudomomentum. The overline denotes an average over one or more full periods / wavelengths. The mid-frequency approximation, i.e. the approximation for GWs with \( N \gg \hat{\omega} \gg f \), then is

\[
\bar{F}_{p,x} = \bar{\rho} < u'w'>
\]

For many waves in the real atmosphere the factor \( 1 - \frac{f^2}{\hat{\omega}^2} \) is very small and accordingly
is neglected in, for instance, all GW parameterization schemes which take into account the effects of GWs in global models (climate models, chemistry-climate models and weather forecast models). In particular, equation (2.20) can be applied directly to model wind data without frequency analysis. This makes zonal means of $F_{px}$ a true reference.

Using the polarization relations (2.19) can be expressed in terms of temperature amplitudes $\hat{T}$ (Ern et al., 2004). Note that for all GWs observable by PREMIEr the difference between equations (6) and (7) of Ern et al. (2004) is less than 1% and can be neglected. We find

$$\hat{F}_{px}, \hat{F}_{py} = \frac{1}{2} \rho \frac{(k_s, l)}{m} \left( \frac{g}{N} \right)^2 \left( \frac{\hat{T}}{\bar{T}} \right)$$

(2.21)

It should be emphasized that (2.21) refers to pseudo momentum flux though it does not contain the factor $1 - \frac{f^2}{\hat{\omega}^2}$. The latter cancels while replacing winds by temperatures based on the dispersion relations. It should be further noted that (2.21) is based on the approximation

$$\frac{\hat{T}}{\bar{T}} = \frac{\hat{\theta}}{\bar{\theta}}$$

(2.22)

where $\theta$ is the potential temperature. Computing this accurately (cf. e.g. Ern et al. (2008)) results in a further correction factor $\zeta$

$$\zeta = 1 + \left( \frac{1}{c_s^2} \right) \left( \frac{\gamma - 1}{\gamma + 1} \right) \left( \frac{g}{2sH} \right)^2 + \left( \frac{m}{g} \right)^2$$

(2.23)

where $c_s$ is the speed of sound, $\gamma = c_p/c_v$ is the adiabatic index following from the specific heat coefficients for constant volume and pressure and $H$ is the scale height. The factor $\zeta$ is smaller than but very close to 1, i.e. $F_{px}$ becomes slightly smaller by applying the correction.

### 2.5 Interaction with the background wind

According to equation (2.18) a GW accelerates or decelerates the mean flow when it dissipates or is refracted by the horizontal background wind shear.

A GW can become instable and starts breaking by either too large wind shear between the different phases of the wave or by static instability. Static instability is caused

---

\[^2\text{Note that GW momentum flux is conserved when only the vertical background wind shear is considered. In this approximation, made frequently by e.g. GW parameterization schemes, only GW dissipation (i.e. breaking) results in GW drag.}\]
when the local temperature gradient (background profile plus wave structure) becomes more negative than the adiabatic temperature gradient $\Gamma$ (cf. Fig. 2.5) and we can determine the maximum stable amplitude $\hat{T}_{\text{max}}$:

$$\Gamma < \frac{\partial (T + \hat{T} \sin(mz))}{\partial z} \Rightarrow \hat{T}_{\text{max}} = \frac{N^2 \overline{T}}{g m} = \frac{N^2 \overline{T} \lambda_z}{2 \pi g} \quad (2.24)$$

The saturation amplitude is hence proportional to the vertical wavelength. As we have discussed above the vertical wavelength depends on the intrinsic phase speed: for instance, if the wave approaches a critical level, the vertical wavelength becomes small and the wave breaks.\textsuperscript{3}

\textsuperscript{3}There is an ongoing discussion what happens to a marginally unstable wave as the one depicted in Fig. 2.5. Does the wave exist and further propagates close to saturation level or does the amplitude break down to a fraction of the stable value? For both there is evidence.
Figure 2.5 also illustrates that the amplitude of a wave increases with height: since momentum flux is conserved and density exponentially decreases with height the amplitude exponentially increases with height. As momentum flux depends on $\hat{T}^2$ the scale height of the amplitude increase is twice as large as the scale height of the density decrease (i.e. about 14 km for the amplitude compared to 7 km for the density).

### 2.6 Summary: Gravity Wave Theory

The most important equation of GW dynamics is the dispersion relation. From the dispersion relation we can infer the group velocity and hence the propagation path of a GW and we can infer to which altitude a GW can propagate before it reaches a critical level. The polarization relations give the relation between the different atmospheric variables such as horizontal wind, vertical wind and temperature which are perturbed by the GW. They allow us to calculate for instance the momentum flux, even if we can measure only one of these variables, provided that we can fully characterize the wave e.g. in terms of its 3-D wave vector. The most important characteristic of such a GW is the amplitude, as it governs the momentum, which the wave conveys, and the phase speed, which decides the altitude where the wave reaches a critical level.
3 Observation techniques and analysis tools

3.1 The Gimballed Limb Observer for Radiance Imaging of the Atmosphere

The new air-borne GLORIA (Gimballed Limb Observer for Radiance Imaging of the Atmosphere; Friedl-Vallon et al., 2006; Ungermann et al., 2010b) employs a two-dimensional (2-D) detector to take an unprecedented amount of spatially resolved spectral samples. GLORIA is unique in being able to pan the line-of-sight of the detector, which allows for viewing at the same volume of air from multiple angles while the carrier moves along the flight path. The combination of these two properties allows to perform tomographic three-dimensional (3-D) retrievals. This new limb-imaging technique allows the detection of meso-scale structures in atmospheric temperature as well as in the chemical composition of the upper troposphere/lower stratosphere (UTLS). Further, GLORIA serves as an air-borne precursor for future satellite-borne instruments.

The instrument and its performance are well documented in a special issue of Atmospheric Measurement Techniques. The article of Friedl-Vallon et al. (2014) gives an overview of the whole system and discusses performances on system level. Piesch et al. (2015) review the mechanical and thermal setup of the spectrometer in detail. Olschewski et al. (2013) and Monte et al. (2014) describe the radiometric calibration system and performance. Kleinert et al. (2014) and Guggenmoser et al. (2015) describe the calibration concept, while Kretschmer et al. (2015) provide an overview of the control and communication architecture of GLORIA.

The GLORIA instrument essentially combines a Fourier transform infrared spectrometer with a two-dimensional detector array measuring in the spectral region from 780 cm⁻¹ to 1 400 cm⁻¹ (see Tab. 3.1 for GLORIA instrument characteristics). The 2-D detector array consists of 256 × 256 pixels, out of which only 48 horizontal × 128 vertical pixels are used for practical reasons to provide more than 6 000 simultaneous limb-views with elevation angles ranging from -3.27° (corresponding to a tangent altitude of roughly 4 km at an observer altitude of 15 km) to slightly upwards and with a horizontal field-of-view (FOV) of 4.07°. To increase the signal-to-noise ratio, all pixels of a row are co-added to so-called superpixels.
Table 3.1: Principal technical characteristics of the GLORIA instrument for two exemplary measurement modes. The panning angle may be slightly reduced depending on the aircraft, onto which GLORIA is be mounted. Both spectral resolution and NESR are given for apodized spectra.

The GLORIA instrument has the ability to pan the line-of-sight relative to the flight direction. This is realized by using a gimbal mount that also stabilizes the line-of-sight during measurements to compensate aircraft attitude variations. While conventional air-borne limb sounders usually point at 90° relative to the flight direction, GLORIA is capable of adjusting its line-of-sight between 45° and 135° (as seen from above the carrier with 0° being the flight direction of the carrier). This allows for viewing the same air volume from multiple angles.

Another advantage of the instrument concept is that it allows to adapt the trade-off between spatial and spectral resolution according to the scientific needs. Four operations modes are currently defined, the “dynamics mode”, the “premier mode”, the “panorama mode”, and the “chemistry mode”. During the POLSTRACC campaign, the “dynamics mode” was used for tomographic retrievals as it offers the highest spatial sampling at a spectral resolution fully sufficient to derive temperature and several major trace gases. As an example for the spectral resolution of the dynamics mode, simulated spectra for mid-latitude atmospheric conditions and 12 km tangent altitude are shown in Fig. 3.1. Even at this comparatively coarse spectral resolution, infrared emission features of important atmospheric constituents are clearly visible.

During the POLSTRACC campaign, GLORIA was mounted within the belly-pod of the High Altitude and Long Range Research Aircraft (HALO). This plane is based on the business jet Gulfstream G550 with modifications that allow the mounting of a wide variety of scientific equipment. The maximum speed of HALO is 940 km/h and the maximum cruise altitude is 15.5 km (DLR, 2010).
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Figure 3.1: Simulated radiance spectra for the GLORIA instrument operating in “dynamics mode” calculated with JURASSIC2. Given are three spectra taken from an observer altitude of 15 km and pointing to tangent heights of 6, 10, and 14 km.

3.2 Retrieval

GLORIA measures spectrally resolved infrared radiation (see Fig. 3.1), but we desire information about temperature and trace gas content. Our understanding of physical processes allows us to readily compute the infrared radiation emitted by atmospheric parcels of known composition and temperature. Deducing the composition and temperature from the observed infrared radiation is a so-called inverse problem, which implies several technical difficulties. Especially in combination with instrument errors inherent to real observations, no realistic theoretical atmospheric state may reproduce a given measurement perfectly, and multiple different theoretical states may fit equally well.

The process of deriving atmospheric quantities from the infrared measurements is generally called retrieval. One requires a radiative transport model to compute simulated measurements for given atmospheric states and a mathematical framework for the inversion. We employ here the Python/C++ based Jülich Rapid Spectral Simulation Code Version 2 (JURASSIC2) to simulate radiances. This model was developed to efficiently handle imager instruments and tomographic retrievals. It is based on JURASSIC (Hoffmann, 2006), which was previously used as forward model for the evaluation of several satellite- and air-borne remote sensing experiments (e.g., Hoffmann et al., 2008; Eckermann et al., 2006; Weigel et al., 2010; Kalicinsky et al., 2013; Ungermann et al., 2016). It contains several approaches of varying computational complexity and accuracy for computing radiances, but we employ in this work the fast table based approach based on the emissivity growth approximation (EGA; e.g., Weinreb and Neuendorffer, 1973; Gordley and Russell, 1981). The model implements also an analytical adjoint based on the DCO tool by Lotz et al. (2012), which allows the efficient computation of Jacobians of the model required for the inversion. The
inversion itself is handled by the python package Juelich Tomographic Inversion Library (JUTIL), which implements a set of convex optimisation algorithms for models supplying efficient adjoints.

GLORIA may operate in different modes, depending on scientific objectives. GLORIA thereby supports to trade available spectral resolution against measurement speed. The spectral resolution may be tuned between 0.0625 cm\(^{-1}\) and 0.625 cm\(^{-1}\). Further, the instrument can be pointed towards different azimuth angles on the right-hand side of the aircraft between 45\(^\circ\) and 135\(^\circ\) with respect to the aircraft heading. This allows in conventional operating modes an alignment with respect to predicted atmospheric structures. But it also enables tomographic observations and retrievals. Conventional aircraft limb-sounding stares at a fixed azimuth angle, which enables the retrieval of vertical cross-section-like products. This is still a standard operation mode of GLORIA. However, panning of the instrument allows to probe the same air mass from different angles and opens up the possibility to perform tomographic retrievals to derive the three-dimensional temperature and trace gas fields in the observed air masses.

Fig. 3.2a shows a full-angle tomographic operation mode (FAT) flight and sampling scenario, which was implemented several times during HALO flights, using a hexagonal flight pattern for practical reasons. Fig. 3.2b shows the limited-angle tomographic operating mode (LAT), which is more limited in its capability of access three-dimensional fields, but imposes less restrictions on the flight plan. Both modes are analyzed in detail in Chap. 5.

The conventional processing of GLORIA dynamics mode measurements, using only a fixed azimuth angle, is now operational and makes up our “standard dynamics mode product” provided to the DLR (2010). Tomographic retrievals currently require manual setup and tuning and are thus only provided for the scientifically most interesting situations. The retrieval setup was improved upon the state described by Ungermann et al. (2015) for the evaluation of TACTS/ESMVAL (Transport and Composition in the upper Troposphere and lowermost Stratosphere / Earth System Model Validation) data. The employed microwindows are listed in Tab. 3.2. The major changes are the inclusion for further spectral windows between 777 cm\(^{-1}\) and 790 cm\(^{-1}\) to improve temperature and ozone and between 808 cm\(^{-1}\) and 835 cm\(^{-1}\) to derive ClONO\(_2\) and get a better background. The correction of thermal window emissions in the calibration was improved such that the 830 cm\(^{-1}\) wavenumber range may be included now and only one continuum may be fitted instead of five for the TACTS/ESMVAL retrievals. Further improvements in the non-linearity correction and flat-fielding of the detector allowed a decrease of regularization strength for temperature and ozone (employed values are shown in Tab. 3.4). This is obviously especially helpful for gravity wave reconstruction. For 1-D retrievals, the given correlation lengths are used. Each tomographic retrievals currently requires a dedicated fine-tuning of correlation length. However, all have in common that in addition to the vertical correlation also horizontal correlation in the two other spatial dimensions is assumed. In addition, also the elevation angle
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Figure 3.2: In Panel (a) a simple schematic of the limb sounding geometry is given. The lines-of-sight (LOS) have a parabolic shape due to the Cartesian coordinate system used for this plot. Panel (b) shows the weighting function along three different LOS indicating the contribution of the respective part of the atmosphere to the observed signal. In panels (c) and (d) the principles of limited angle tomography (LAT) and full angle tomography (FAT) with the GLORIA instrument are depicted, respectively. The dots indicate the tangent points and are coloured according to their altitude. Each grey sector indicates one horizontal scan from 45° to 135°. The lighter the grey, the later in time are these measurements taken. In practise, the measurement density is much denser than depicted here.
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<td>956.8750 – 962.5000</td>
</tr>
<tr>
<td>10</td>
<td>794.3750 – 795.0000</td>
<td>30</td>
<td>980.0000 – 984.3750</td>
</tr>
<tr>
<td>11</td>
<td>795.6250 – 796.2500</td>
<td>31</td>
<td>992.5000 – 997.5000</td>
</tr>
<tr>
<td>12</td>
<td>796.8750 – 797.5000</td>
<td>32</td>
<td>1000.6250 – 1006.2500</td>
</tr>
<tr>
<td>13</td>
<td>798.1250 – 798.7500</td>
<td>33</td>
<td>1010.0000 – 1014.3750</td>
</tr>
<tr>
<td>14</td>
<td>799.3750 – 799.3750</td>
<td>34</td>
<td>1388.1250 – 1389.3750</td>
</tr>
<tr>
<td>15</td>
<td>808.1250 – 808.1250</td>
<td>35</td>
<td>1390.0000 – 1391.2500</td>
</tr>
<tr>
<td>16</td>
<td>808.7500 – 808.7500</td>
<td>36</td>
<td>1391.8750 – 1393.1250</td>
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<td>809.3750 – 809.3750</td>
<td>37</td>
<td>1393.7500 – 1395.0000</td>
</tr>
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<td>820.6250 – 820.6250</td>
<td>38</td>
<td>1395.6250 – 1396.8750</td>
</tr>
<tr>
<td>19</td>
<td>821.2500 – 821.2500</td>
<td>39</td>
<td>1397.5000 – 1398.7500</td>
</tr>
</tbody>
</table>

Table 3.2: A list of integrated spectral windows (ISW) employed for the conventional dynamics mode retrieval and their spectral range.

<table>
<thead>
<tr>
<th>#</th>
<th>spectral range (cm⁻¹)</th>
<th>#</th>
<th>spectral range (cm⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>790.6250 – 791.2500</td>
<td>7</td>
<td>918.7500 – 923.1250</td>
</tr>
<tr>
<td>1</td>
<td>791.8750 – 792.5000</td>
<td>8</td>
<td>956.8750 – 962.5000</td>
</tr>
<tr>
<td>2</td>
<td>793.1250 – 795.0000</td>
<td>9</td>
<td>980.0000 – 984.3750</td>
</tr>
<tr>
<td>3</td>
<td>796.8750 – 799.3750</td>
<td>10</td>
<td>992.5000 – 997.5000</td>
</tr>
<tr>
<td>4</td>
<td>883.7500 – 888.1250</td>
<td>11</td>
<td>1000.6250 – 1006.2500</td>
</tr>
<tr>
<td>5</td>
<td>892.5000 – 896.2500</td>
<td>12</td>
<td>1010.0000 – 1014.3750</td>
</tr>
<tr>
<td>6</td>
<td>900.0000 – 903.1250</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 3.3: A list of integrated spectral windows (ISW) employed for the tomographic retrieval and their spectral range.
Observation techniques and analysis tools

<table>
<thead>
<tr>
<th>parameter</th>
<th>value</th>
<th>parameter</th>
<th>value</th>
</tr>
</thead>
<tbody>
<tr>
<td>(c_{\text{aerosol}})</td>
<td>640 km</td>
<td>(c_{\text{temperature}})</td>
<td>0.3 km</td>
</tr>
<tr>
<td>(c_{\text{CCl}_4})</td>
<td>2 km</td>
<td>(c_{\text{CClONO}_2})</td>
<td>2 km</td>
</tr>
<tr>
<td>(c_{\text{CFC-11}})</td>
<td>8 km</td>
<td>(c_{\text{CFC-12}})</td>
<td>8 km</td>
</tr>
<tr>
<td>(c_{\text{H}_2\text{O}})</td>
<td>20 km</td>
<td>(c_{\text{HNO}_3})</td>
<td>0.4 km</td>
</tr>
<tr>
<td>(c_{\text{O}_3})</td>
<td>8 km</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 3.4: Vertical correlation lengths employed for the regularization.

of each measurement is retrieved to correct for known deficiencies of the elevation angle stabilization system.

For the tomographic processing, a reduced set of microwindows is employed, as we are mainly interested in the parameter temperature for gravity wave reconstruction. This allows the discarding of microwindows for non-relevant trace gases such as water vapor, CFCs, or ClONO\(_2\). This drastically reduces the required computational resources for the tomographic retrievals and allows to perform sensitivity studies in contrast to only individual retrievals. The remaining windows capture CO\(_2\), O\(_3\), and HNO\(_3\) emissions as well as a rough reconstruction of some interfering trace gases.

The errors of the retrieved quantities are analyzed using a linear approximation (Rodgers, 2000). The retrieval result is effectively the sum of the true atmospheric state smoothed by the averaging kernel matrix \(A\), the a priori influence, and measurement errors. The averaging kernel matrix is thereby the product of the so-called gain matrix \(G\) and the Jacobian matrix of the forward model (Ungermann et al., 2015). The gain matrix can be thought of as a (regularised) pseudo-inverse of the Jacobian matrix from the forward model.

Given a covariance matrix \(S\) describing the effect of an arbitrary error source on the measurements, the gain matrix \(G\) can be used to linearly estimate a covariance matrix describing the effect of this error source on the retrieval result as \(G^TSG\). Such a covariance matrix can be readily assembled at least approximately for many systematic error sources using standard deviations and a reasonable (potentially multi-dimensional) vertical correlation length and involving an auto-regressive approach (Tarantola, 2004). We distinguish between random errors resulting from measurement noise and other, usually systematic, error sources. The measurement noise is taken from theoretical estimates given by Friedl-Vallon et al. (2014) that agree well with estimates derived from measurements (Kleinert et al., 2014). For integrated spectral windows covering several samples, the assumed noise is divided by the square root of the number of spectral samples. In addition, a fixed relative noise component of 0.1 % is assumed for all integrated spectral windows.

The error estimate associated with the noise error source is given as precision value. The errors resulting from misrepresented background gases, uncertainties in spectral
line characterization (taken to be 5% under the assumption that, statistically, some errors in individual line parameters cancel each other out), uncertainties in instrument attitude, and calibration errors are combined under the label accuracy. It is assumed that gain and offset errors are spatially uncorrelated, but spectrally fully correlated (in the absence of a better characterization, this provides a worse error estimate than assuming no spectral correlation).

The sum over each row of the averaging kernel matrix $A$ is supplied as measurement contribution. The full width at half maximum of each row is also computed using linear interpolation to provide a measure of the vertical resolution. The smoothing error is not given, as the underlying covariance matrix $S_a$ describing the actual statistics of the fine structure of the real atmospheric state is far from being known accurately in an optimal estimation sense. Still, the vertical resolution and measurement contribution can be used to gain insight into the quality of the data. For tomographic retrievals also the horizontal resolution is determined from the major axes of an ellipsoid fitted to the convex hull of all points contributing more than half the maximum value of the averaging kernel matrix row.

### 3.3 Nadir sounding: AIRS

The Atmospheric Infrared Sounder (AIRS) is one of the instruments onboard NASA's Earth Observing System (EOS) Aqua satellite (see also Aumann et al., 2003; Chahine et al., 2006). AIRS is a nadir-scanning instrument that performs scans across the satellite track. Each scan consists of 90 footprints across track, and the width of the swath is about 1800 km. At nadir, the footprint diameter is 13.5 km, and the across-track sampling step is 13 km. The along-track sampling distance is 18 km. The EOS Aqua satellite is in a sun-synchronous orbit with the fixed equator crossing times of 13:30 local time for the ascending orbit (flying northward) and 01:30 local time for the descending orbit (flying southward).

AIRS is a hyperspectral radiometer that measures atmospheric emissions of CO$_2$ with high spectral resolution. In contrast to limb geometry, nadir sounding depends on the saturation of the radiance along the ray-path. Depending on the wavelength, lines or bands of different strength of the absorption cross section are observed. For the spectral band of CO$_2$ emissions this is visualized in Fig. 3.3 in terms of vertical weighting functions for AIRS. The altitude resolution depends entirely on these weighting functions. By combining different IR wavelengths, weighting functions centered at different altitudes can be used and a retrieval of an altitude profile can be performed. In that sense the altitude resolution would depend on the spectral resolution and the number of independent pieces of information in terms of different center altitudes of these weighting functions.

However, the response to a wave may be worse than the altitude resolution. We can
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understand this complication by a thought experiment. Let us assume for simplicity that the weighting function takes the shape of a boxcar filter with a width of 10 km. If we shift that boxcar with a fine increment over a step signal, we get a ramp and can infer the original step. However, if we apply the boxcar to a 10 km sinusoid, the net result is zero and nothing can be retrieved.

The AIRS operational temperature retrieval (cf. Susskind et al., 2003; Cho and Staelin, 2006) combines 9 AIRS footprints for a cloud-clearing procedure. Therefore, a special temperature retrieval optimized for the analysis of gravity waves was developed by Hoffmann and Alexander (2009). In this retrieval, temperatures are derived separately for each AIRS footprint, thereby preserving the full horizontal resolution of AIRS. For daytime data, temperatures are derived from the 15 µm emission of CO₂. For retrievals of nighttime data, emissions of the 4.3 µm and the 15 µm CO₂ can be combined. This is the case because during nighttime the 4.3 µm emission of CO₂ is not affected by non-local thermodynamic equilibrium (non-LTE) effects. Correspondingly, AIRS nighttime data are sensitive to GWs of somewhat shorter vertical wavelengths than in the case of daytime data. In addition, the noise of nighttime data is somewhat reduced. Except for polar latitudes, daytime data correspond to ascending orbits, and nighttime data to descending orbits, respectively.
The sensitivity of the nighttime retrieval for gravity waves of a given vertical wavelength was investigated by Hoffmann and Alexander (2009). The effective vertical resolution that is given by the retrieval averaging kernels varies between 7 km at 20 km altitude and 15 km at 60 km altitude. The AIRS sensitivity function for the middle stratosphere is given in Fig. 3.4 for amplitudes (blue curve) and squared amplitudes or momentum fluxes (red curve). These values do not include effects of background removal and cases of GWs with horizontal wavelengths < 100 km that may be affected by the limited AIRS footprint size. Sensitivity functions that also include the effect of background removal by an across-track 4th-order polynomial (a standard procedure of background removal for nadir sounders) are given, for example, in Meyer et al. (2018) or the supporting information of Ern et al. (2017).

From Fig. 3.4 we can see that the sensitivity vanishes for any waves with vertical wavelengths below 10 km. For waves larger than 15 km, reasonable responses are found. Unfortunately, this removes a major part of the momentum flux. Applying a 12 km vertical wavelength cut-off to gravity wave momentum flux (GWMF) resolved in ECMWF reduces the momentum flux by almost an order of magnitude, as is shown in Fig. 3.5 (note the logarithmic color scale). This graphic does not include the effect that sensitivities are much less than 1 for vertical wavelengths between 12 and 20 km and the additional influence of background noise. In the end, only the strongest gravity wave events will be identified unambiguously by a nadir viewing satellite. Please note that these calculations were performed for AIRS, but are based on the fundamental principles of nadir temperature measurements, thus also apply to new nadir sounders for instance in the Sentinel program.

So far we have discussed the limits of nadir sounders. The positive facts are: the
horizontal resolution depends mainly on the size of the footprint. In addition, the general very wide swath width allows to assess a wave field in its whole.

**Pros and cons of nadir sounders**

In summary, nadir sounders miss most of the total gravity wave momentum flux (10% or less visible) as they do not see gravity waves of vertical wavelengths of 10 km and shorter. In particular, they miss almost all of those waves which are actually interacting with the background flow by breaking. Nadir sounders are therefore not suited for investigations of the momentum balance.

A vertical wavelength threshold of 12 km corresponds to an intrinsic phase speed of 40 m s\(^{-1}\) or faster. This means either high background winds or a source process which generates fast waves are required for waves being visible to nadir sounders. In cases of Doppler shift, the waves propagating opposite to the wind are refracted to the longest wavelengths. The visibility limit of nadir sounders hence also could influence the direction characteristic derived for the waves observed.

On the positive side, from nadir data the horizontal propagation direction of gravity waves can be determined and, with some uncertainty, also the vertical wavelength. As has been shown by Ern et al. (2017) or Wright et al. (2017) gravity wave momentum flux can be estimated from nadir sounders. Nadir sounders can therefore be employed for case studies of gravity wave momentum flux. Also, models can be validated if a) an observational filter is applied and b) there are sufficient waves in the visible phase speed range, i.e. the model does not release all gravity wave momentum flux in the invisible range.

For aircraft campaigns, the limited sensitivity should be taken into account in the flight planning. If this is done, the potential to find again a gravity wave identified at
lower altitudes after propagation to a higher altitude may be demonstrated. This will be discussed below.

### 3.4 Description of ECMWF analysis data

Modern numerical weather prediction (NWP) relies on two fundamental components, first, a high-resolution general circulation model (GCM) which includes all processes relevant in the atmosphere and, second, the assimilation of a multitude of different types of measurements in order to represent as accurately as possible the current state of the atmosphere. For this study we use analysis data of the European Centre for Medium range Weather Forecast (ECMWF). The ECMWF integrated forecast system assimilates data by 4D var. This constrains the synoptic scale situation up to an altitude of \( \sim 40 \text{ km} \) and provides GWs a close-to-reality background for propagation. Also excitation due to orography and spontaneous emission are well described. The assimilation does not constrain the GWs itself, thus, they can develop freely from the model physics.

The dynamical core of the ECMWF GCM is based on a spectral representation of the atmosphere. The spatial resolution was enhanced several times in the recent decade. The earliest data we use (starting from January 2006) are generated with a model resolution N400/T799 corresponding to a spatial grid of about 25 km. In the vertical 91 levels were used, which are densely spaced close to the surface and become sparse (several kilometer spacing) towards the model top. At 26 January 2010 the horizontal resolution was increased to N640/T1279 which corresponds to about 16km. On 25 July 2013 the vertical resolution was increased to 137 levels. The larger number of vertical levels is partly used to extend the model top, partly to increase the resolution in the stratosphere. Though the dynamical core would in principal allow to resolve waves of 50 km (30 km after 2010), hyperdiffusion, which was introduced to provide numerical stability, limits well-resolved waves to about 10 spatial grid points (Preusse et al., 2014). An example how the true resolution can be estimated is demonstrated in Fig. 3.6. According to theory, the power spectral density should follow a slope of \(-5/3\) (purple line). However, the power spectral density of the model (red line) drops at some point. This dropping is associated with the start of the hyperdiffusion. Thus, we may assume that between 2006 and 2010 waves of horizontal wavelengths longer than \( \sim 250 \text{ km} \) are fully resolved, and afterwards waves of horizontal wavelengths longer than \( \sim 160 \text{ km} \). Shorter waves, if excited e.g. by topography, may still be present but are suppressed in amplitude.
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Figure 3.6: Power spectra of ECMWF temperatures in zonal direction averaged over the period from 28 January to 3 February 2008 and over latitudes from 40° to 60° N. Altitude is 25 km. The red line shows the average power spectral density, the purple line indicates a slope of $-5/3$, the green line indicates the fit-by-eye where the spectrum becomes significantly steeper than the power-law, corresponding to a wavelength of $\sim 220$ km. Vertical blue lines are drawn for horizontal wavelengths of 10, 100, 1000 and 10,000 km. From Preusse et al. (2014).

3.5 Background removal

The atmosphere is shaped by dynamical features of different spatial and temporal scales. Hence, the first step for detailed gravity wave analysis of any data set is the separation of the GW fluctuations from a general background atmosphere. The stratosphere is dynamically dominated by the general zonal mean state of the atmosphere, synoptic planetary waves and small-scale gravity waves. By subtracting the global wave modes, the remaining fluctuations can be assumed to be gravity waves. In the spherical earth system, the global wave modes have integer zonal wavenumbers and vary slowly in latitude and altitude. In the mid to higher stratosphere, synoptic and small-scale wave signals separate clearly in spectral space, such that background and gravity wave fluctuations are traditionally separated using zonal Fast Fourier Transform (FFT) highpass
filtering with a cut-off wavenumber of 6.

Wave fluctuations, also those introduced by synoptic waves, grow with altitude. Hence, the separation is more difficult at lower altitudes, where synoptic effects can influence higher wavenumbers than at higher altitudes. In the lower stratosphere, a filtering of wavenumbers $\leq 12$ is usually still enough to erase those synoptic effects from the temperature fluctuation field associated with gravity waves. In the troposphere and for wind fluctuations, filtering up to even higher wavenumbers can be applied to extract the wave fluctuations. However, the synoptic and gravity wave signals may overlap in the spectrum. In addition, polynomial smoothing with a Savitzky-Golay filter in the vertical and/or latitudinal direction can be applied to smear out remaining wave signals from a background atmosphere. Systematic tests of the background removal for different atmospheric parameter fields using different cut-off wavenumbers for ECMWF model data are presented in Section 9.2.1.

The spectral filtering approach works well as long as the data set is global, since the Fast Fourier Transformation assumes spatial periodicity in zonal direction. In order to remove the background self-consistently from local GLORIA data, a local background removal is necessary. To identify gravity waves in the retrieved 3-D temperature field, this was done by applying polynomial smoothing with a Savitzky-Golay filter of third-order over a chosen spatial distance, e.g. 3.125 km, 750 km, 750 km, in vertical, zonal, and meridional direction. The values of the calculated polynomials at the respective points are treated as background atmosphere. Subtracting the polynomial values from the original data then results into a local fluctuation field.

Single profile observations or observations along a flight leg do not provide three-dimensional information. A background removal in such cases has to be calculated from two-dimensional fields. A common approach is again smoothing, this time along the profile or flight track.

### 3.6 Three-dimensional sinusoidal fitting (S3D)

The S3D method – a method to fit a sinusoid to perturbation data – has been developed at IEK-7 in the last years and has been used for wave characterisation in several publications since (Lehmann et al., 2012; Preusse et al., 2014; Ern et al., 2017; Krisch et al., 2017). The method can in principle be applied to identify gravity wave patterns in any three-dimensional perturbation data set. At IEK-7, we usually use temperature data from different observation and model systems, like GLORIA retrievals, satellite retrievals and analysis fields from the ECMWF Integrated Forecasting System (IFS).

The S3D method fits a sinusoid $s(k, x) = \hat{T} \sin(kx + ly + mz + \phi)$ to the temperature
perturbations $T'$ by minimizing the model error

$$\chi^2 = \sum_i \frac{T'-s(k,x)}{\sigma^2}$$

within an analysis cube. Here $x = (x, y, z)$ denotes the location vector and $k = (k, l, m)$ the wave vector with $k, l, m$ the wavenumbers in $x, y, z$-directions. The wavenumbers are calculated in an iterative minimization scheme reducing $\chi^2$. We have different minimization approaches implemented, such as a parameter nesting and a steepest descent method. The temperature amplitudes $\hat{T}$ and phase $\phi$ can be calculated analytically in each iteration step from the fitted wavenumbers. The iteration is ended when the method converges (i.e. $\chi^2$ falls below a limit) or a maximum number of iteration steps is reached.

The method works on analysis cubes – chosen three-dimensional sub-regions of the perturbation field. In each cube, a superposition of monochromatic sine waves is assumed to allow for the fitting. The method is dependent on the cube size used for the fit assuming constant sampling distances. If the cube is too large compared to the wavelengths, the method looses information about small fluctuation that gets masked by larger scales. If the cube is too small, the phase gradients and hence the wavelengths cannot be determined with sufficient accuracy in the presence of variations, noise etc. Tests have shown that the wavelength of the wave can be determined with reasonable accuracy, if the ratio of wavelength to cube size is less than 2.5.

Localization weights help to determine the amplitudes and phases with more accuracy. A gaussian weight reduces the impact of data distant from the cube center. When using large analysis cubes compared to the signal wavelengths, this reduces errors e.g. arising from growing wavelengths within the analysis cube. The corresponding scale factor in the model is calculated as

$$\frac{1}{\sigma^2} = e^{-\frac{(x-x_c)^2}{\psi_x^2}} e^{-\frac{(y-y_c)^2}{\psi_y^2}} e^{-\frac{(z-z_c)^2}{\psi_z^2}}.$$ (3.2)

To interpret GLORIA retrieval results with regard to gravity waves, a weighting function $\sigma^2(x)$ for the tangent point density is used. The impact of data with a low tangent point density is reduced by choosing the weight to be 1 if a tangent point exists in the corresponding grid cell and $10^5$ if not. Systematic tests of this algorithm by superposition of two sinusoids show that even for wavelengths up to 2.5 times the cube size, the wave parameters of both waves are fitted with errors below 1%.

We can calculate additional parameters like wavelengths, propagation direction and speed, intrinsic wave frequency and pseudo momentum flux which are characterizing the wave. More details about important wave parameters and their calculation have been addressed in Chapter [2].
3.7 The the Gravity Wave Regional Or Global RAy Tracer (GROGRAT)

The ray tracing tool used to predict gravity wave propagation is the Gravity Wave Regional or Global Ray Tracer (GROGRAT). GROGRAT is a well established model to trace the propagation of gravity waves in the atmosphere (Marks and Eckermann, 1995). Based on the WKB assumption, it derives the group velocity of a gravity wave from the dispersion relation. This allows to estimate the location of a gravity wave packet after a time step $\delta t$. The new location is associated with a different background state (winds, buoyancy frequency) of a given atmosphere. This induces changes of the wave vector $k$. Together, location vector $x$ and wave vector $k$ form the state vector of the gravity packet which is projected in time according to the ray-tracing equations (Lighthill, 1978; Marks and Eckermann, 1995) using a fourth-order Runge-Kutta scheme. The ray-tracer can be run backward in time in order to identify the origin of the wave. In this way, wave excitation by potential gravity wave source processes can be investigated. Alternatively, GROGRAT can be run forward in time in order to study the further propagation of a wave packet.

For forward propagation, the wave amplitude and momentum flux are calculated based on the conservation of wave action flux along the ray. GROGRAT takes gravity wave saturation (cf. Chapter 2) due to statical and dynamical instability into account as well as gravity wave damping due to radiative and turbulent dissipation.

In order to run GROGRAT, the full state vector of a wave at the launch location and time and a background atmosphere are required. The launch values can be provided by the S3D fitting results either of GLORIA, ECMWF or AIRS data. For the background we use low-pass filtered ECMWF data applying the same Fast Fourier Transformation filtering approach as used for the S3D background removal.

3.8 The Mission Support System (MSS)

The Mission Support System (MSS) is a web-service based tool for flight planning. MSS was initially develop by Rautenhaus et al. (2012) at DLR for mission support of HALO campaigns. In the preparation of the GWEX campaign, we arranged for MSS to enter the public domain and become an open source program. In this way, we could fix problems and implement needed features for the current as well future campaigns and open up development also for third parties.

1Technically, GROGRAT uses the position in 3-D space, the horizontal wave vector, the ground-based frequency and the wind amplitude as launch parameters. From the dispersion relation, both the ground-based frequency and the wind amplitude can be calculated from the wave vector and temperature amplitudes.
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Figure 3.7: An exemplary screenshot of MSS planning of a gravity wave specific flight over Iceland. On the left-hand side a top view with a horizontal cross-section through our gravity-wave forecast product at 11 km is shown. The remote sensing tool is opened, depicting tangent points in red and flight sections with problematic solar interference in red colors. The right-hand side shows a vertical cross-section and a part of the tabular flight plan view.

The MSS is a Python-based tool consisting of two major components. First, a web-server providing an Open Geospatial Consortium Web Map Service (WMS) with some additional proprietary features enabling, for example, also vertical cross-sections needed for flight planning. Second, a cross-platform Python client for actual flight planning, which may access any WMS service. Python is a well suited choice here, as web programming is well supported, it can operate on any operating system, and the effort for implementing new features is very low. Even MSS users without deep Python experience were able to quickly implement new features.

The WMS server is able to provide horizontal and vertical cuts through any regularly gridded model data following the NetCDF Climate and Forecast metadata standards. Some work was done on our side to make the setup of this component much simpler for such standardized NetCDF data as well as by implementing plots of new data products such as gravity wave residuals or momentum flux data. It is now straightforward to include new models and new data products with standard plots into the WMS server.

The frontend was drastically enhanced, mostly by fixing long-standing bugs such as that no images could be saved to disk. In addition, several features required for flight planning for GLORIA were added, such as a display of tangent points of mea-
surements, inclusion of solar azimuth angles for avoiding direct sunlight on the optics. Figure 3.7 shows an exemplary arrangement of MSS windows for a flight planning of a tomographic measurement flight over Iceland.

Further, the program was ported to the Anaconda Python package management platform. Whereas previously the dependencies needed to be installed manually, a process complicated to the point that most people did not succeed in installing the frontend, it can now be installed by a simple double click. This enabled the use of the program by all participants of the flight planning meetings, leading to more inputs and overall better flight plans. A full list of changes can be seen in the change logs of the last software releases (https://bitbucket.org/wxmetvis/mss).

3.9 ICON-ART simulations

ICON-ART is the combination of the global non-hydrostatic forecast model ICON (ICOsa-hedral Nonhydrostatic) with the extension ART (Aerosols and Reactive Trace gases). ICON is the new generation meteorological forecast system at the German Weather Services (DWD, Deutscher Wetterdienst, Offenbach Germany) and is developed in cooperation with Max-Planck-Institute for Meteorology, Hamburg (Zängl et al., 2015). To simulate influences of chemistry and aerosols, the module ART is being developed by the KIT (Schröter, 2018; Schröter et al., 2018). The model description summarized in the following and the model parameterization used here in particular are described in detail by Diekmann (2017).

ART couples the temporal and spatial state of trace gases and aerosols to each model time step based on the actual atmospheric state. The influence of clouds and radiation is furthermore coupled back to the meteorology at the same time. Tracer simulations by ART are based on the transport scheme of ICON. Thereby, transport by horizontal and vertical advection as well as turbulence are simulated by solving the continuity solution. ART simulates production and destruction of reactive trace gases and is capable of considering trace gas emission scenarios.

In contrast to other forecast systems based on regular latitude-longitude grids, ICON involves a triangular grid. The triangular structure fits better with modern computer architectures, and the dynamical core scales very well. The choice of a triangular grid avoids a meridional convergence and singularities at the poles, improves mass conservation and allows local grid refinement (nesting). Here, we use ICON-ART simulations based on the R2B06 grid with an effective resolution of 39.5 km. For analyses, the data on the native triangular ICON grid are resampled to a regular 0.5° x 0.5° latitude-longitude grid based on the nearest-neighbor method. Considerably higher resolutions are being established for future simulations.

The vertical discretization is based on a hybrid height coordinate according to Leuenberger et al. (2010). At the ground the coordinate follows the local topography. Above,
the coordinate continuously turns to constant height levels, which are approached at 16 km altitude. Here, a configuration involving 90 vertical levels between ground and 75 km is used, with a vertical resolution of about 400 m in the considered altitude range.

To investigate mountain wave signatures in trace gas fields, we use an ICON-ART simulation for the entire polar winter 2015/16 by Diekmann (2017). The simulation is based on the so-called Poor Mans Nudging forecast mode: the dynamical core of ICON is updated with the actual state of the atmosphere in defined time intervals (here: every day at 0 UTC). Thereby, the meteorological situation (e.g. pressure, temperature, potential vorticity, specific humidity) is updated by operational ECMWF Integrated Forecast System (IFS) data at T1279L60 resolution, resampled to the native ICON grid. Therefore, the meteorological variables in this setup show small discontinuities at the reinitialization points, while the trace gas fields simulated by ART are simulated continuously.

Here, we analyze wave signatures modeled by ICON-ART using water vapor based on specific humidity (qv) and ozone. While specific humidity is subject to the Poor Mans Nudging, ozone is simulated continuously over the entire winter. Specifically, ozone loss is simulated based on a cold tracer (Braesicke and Pyle, 2003). The cold tracer is an indicator for air masses suitable for the existence of polar stratospheric clouds and capable of heterogeneous chlorine activation. The cold tracer is activated if the simulated temperatures fall below a threshold of 195 K and declines exponentially at warmer temperatures to account for deactivation of ozone-depleting species. Here, ozone depletion was simulated aided by a cold tracer with a life time of 2 days.

The work by Diekmann (2017) has shown that ICON-ART trace gas simulations reproduce GLORIA observations in a realistic way. Therefore, ICON-ART provides the prerequisites to analyze mountain wave-induced signatures in tracer fields.
4 Gravity Wave Observation Probability

In preparation of the campaign, we investigated the likelihood for GW events suitable for the aims of GWEX. We based this on the results of the GLORIA sensitivity study performed in GWEX Task 2 (see Chap. ) and deduced minimum criteria required for a GW to be observable by GLORIA. A second aim was that the GW should be able to propagate into the stratosphere and be also observable to AIRS. Together this leads to the criteria listed in Tab. 4.1. An additional criterion is the ground-based period of GWs, which must be substantially larger than the time required to perform the tomographic flight pattern required by GLORIA.

<table>
<thead>
<tr>
<th>instrument</th>
<th>reference altitude</th>
<th>condition</th>
</tr>
</thead>
<tbody>
<tr>
<td>GLORIA</td>
<td>12 km</td>
<td>$\lambda_z \leq 12$ km</td>
</tr>
<tr>
<td>GLORIA</td>
<td>12 km</td>
<td>temp. amplitude $\geq 1.5$ K</td>
</tr>
<tr>
<td>AIRS</td>
<td>40 km</td>
<td>$\lambda_z \geq 10$ km</td>
</tr>
<tr>
<td>AIRS</td>
<td>40 km</td>
<td>temp. amplitude $\geq 5$ K</td>
</tr>
</tbody>
</table>

Table 4.1: Simplified criteria for events measurable by GLORIA and AIRS

Based on these criteria we now reviewed the GWs occurring in January for 10 successive years from 2006 to 2015. For this purpose, we ran S3D analysis for an altitude of 12 km for ECMWF analysis data with the best available resolution for this individual year. The analyses results were compared to the GLORIA observation criteria. The S3D results were also used to project the wave to an altitude of 40 km, where the so obtained GW parameters were compared to the AIRS criteria. This method ensures that we investigate the AIRS-observability for GLORIA-observable events. ECMWF data are available in 6-hourly time steps. For each of these 124 time steps of a complete January a map as displayed in Fig. 4.1 was produced. The red dots show the events of interest that can be observed by both GLORIA and AIRS.

Viewing these in total 1240 maps and using some experience in GW physics, we selected a list of events of interest for these whole 10 years. These events are listed in Tabs. 4.2, 4.3, and 4.4. The list shows for very year four or more cases where the criteria are matched. The tables also allows to select priority target regions, where events
Figure 4.1: Example of a map displaying the simplified criteria for an event of interest at 9 January 2014, 06 GMT. Blue dots show all events where the vertical wavelength is shorter than 12 km and the amplitude is larger than the threshold $T_G$ of 1.5 K (cf. picture header). Red dots show those points where, in addition, also the AIRS criteria are matched, that is a minimum vertical wavelength of 10 km and a minimum amplitude $T_A$ of 5 K. In the example shown there is a region of interest north of Iceland.

are most likely. These are Northern Scandinavia, Southern Scandinavia, Iceland and Greenland. In the end, we found during the campaign GW events in all these regions, except for only weak activity at northern Scandinavia. The two GWEX flights were performed over Iceland and Southern Scandinavia. In order to understand why we did not investigated GWs close to our campaign base in Kiruna, we have to set the winter 2016 in relation to the typical long-year average winter. The actual campaign phase in Kiruna started on 10 January 2016. Accordingly, Fig. 4.2 compares for the time period 10 January to 31 January the winds from a 10-year average with the ones for 2016. The winter 2016 had record cold polar cap temperatures in the stratosphere. This had several consequences. In the stratosphere the polar vortex was, on average, very strong. In general, the high-wind regions were shifted to the south. And, at the lowest altitude, wind speeds were lower over northern Scandinavia. In particular the latter fact explains, why we had less likelihood than usual to find GWs close to the base Kiruna in northern Scandinavia.
Figure 4.2: Comparison of the average winds from ERA-interim for the time period 10 January to 31 January. The left column shows a 10-year average, the right column the year 2016. The rows show altitudes of 36 km, 27 km and 11.5 km.
### Table 4.2: List of prospective cases identified via the survey of 10 years of January by criterion maps: 2006-2010

<table>
<thead>
<tr>
<th>GMT</th>
<th>location</th>
<th>likely source</th>
<th>comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>2006-01-07T18:00Z</td>
<td>S.Green</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2006-01-09T12:00Z</td>
<td>Norway</td>
<td>jet</td>
<td>well observable, masked by synoptic</td>
</tr>
<tr>
<td>2006-01-11T18:00Z</td>
<td>Iceland, Scand</td>
<td>orography</td>
<td>well observable, moderate amplitude</td>
</tr>
<tr>
<td>2006-01-18T00:00Z</td>
<td>Iceland</td>
<td>jet inst.</td>
<td>short $\tau_{gb}$, linear flight?</td>
</tr>
<tr>
<td>2007-01-06T12:00Z</td>
<td>Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2007-01-09T18:00Z</td>
<td>Scand, C.Green</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2007-01-11T12:00Z</td>
<td>S.Green</td>
<td>oro+jet</td>
<td>well observable, long train, linear+hex?</td>
</tr>
<tr>
<td>2007-01-16T00:00Z</td>
<td>Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2007-01-22T12:00Z</td>
<td>C.Green</td>
<td>orography</td>
<td>reasonably well observable</td>
</tr>
<tr>
<td>2007-01-24T12:00Z</td>
<td>C.Green</td>
<td>orography</td>
<td>well observable, shipwave, wind $\parallel$ coast</td>
</tr>
<tr>
<td>2007-01-29T06:00Z</td>
<td>Norway</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2008-01-01T06:00Z</td>
<td>S.Green</td>
<td>front?</td>
<td>well observable</td>
</tr>
<tr>
<td>2008-01-19T00:00Z</td>
<td>S.Green</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2008-01-20T00:00Z</td>
<td>S.Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2008-01-25T18:00Z</td>
<td>S.Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2008-01-29T00:00Z</td>
<td>S.Green, Scand</td>
<td>orography</td>
<td>well observable, cf. Preusse et al. (2014)</td>
</tr>
<tr>
<td>2009-01-05T12:00Z</td>
<td>N.Green, N.Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2009-01-06T18:00Z</td>
<td>N.Green, N.Scand</td>
<td>oro+jet</td>
<td>well observable</td>
</tr>
<tr>
<td>2009-01-11T00:00Z</td>
<td>C.Scand</td>
<td>oro+jet</td>
<td>well observable</td>
</tr>
<tr>
<td>2009-01-15T18:00Z</td>
<td>N.Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2010-01-01T18:00Z</td>
<td>C.Green</td>
<td>front?</td>
<td>long-drawn wave fronts, use linear flight</td>
</tr>
<tr>
<td>2010-01-02T06:00Z</td>
<td>N.Green, Atlant.</td>
<td>oro+jet</td>
<td>use linear flight?</td>
</tr>
<tr>
<td>2010-01-10T12:00Z</td>
<td>C.Green, N.Scand</td>
<td>orography</td>
<td>well observable, Green: shipwave, wind $\parallel$ coast</td>
</tr>
<tr>
<td>2010-01-27T18:00Z</td>
<td>C.Green</td>
<td>oro+jet</td>
<td>well observable</td>
</tr>
<tr>
<td>GMT</td>
<td>location</td>
<td>likely source</td>
<td>comment</td>
</tr>
<tr>
<td>----------------</td>
<td>--------------</td>
<td>---------------</td>
<td>----------------------------------------------</td>
</tr>
<tr>
<td>2011-01-02T00:00Z</td>
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<td>front</td>
<td>well observable</td>
</tr>
<tr>
<td>2011-01-03T06:00Z</td>
<td>C.Green</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2011-01-09T00:00Z</td>
<td>C.Green</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2011-01-10T18:00Z</td>
<td>S.Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2011-01-16T18:00Z</td>
<td>S.Scand</td>
<td>orography</td>
<td>observable, $\lambda_h \approx 500$ km</td>
</tr>
<tr>
<td>2011-01-19T06:00Z</td>
<td>Iceland</td>
<td>jet-exit</td>
<td>well observable</td>
</tr>
<tr>
<td>2011-01-23T12:00Z</td>
<td>Iceland</td>
<td>jet</td>
<td>well observable</td>
</tr>
<tr>
<td>2011-01-25T18:00Z</td>
<td>C.Green</td>
<td>front+oro</td>
<td>well observable</td>
</tr>
<tr>
<td>2011-01-27T18:00Z</td>
<td>N.Scand</td>
<td>orography</td>
<td>well observable, waves all along Scand. coast</td>
</tr>
<tr>
<td>2011-01-28T18:00Z</td>
<td>C.Green, N.Scand</td>
<td>shear / oro</td>
<td>well observable, Green: large displacements ($v+h$)</td>
</tr>
<tr>
<td>2012-01-09T12:00Z</td>
<td>S.Green</td>
<td>oro(+jet?)</td>
<td>well observable</td>
</tr>
<tr>
<td>2012-01-10T18:00Z</td>
<td>C.Scand, S.Green</td>
<td>oro+jet</td>
<td>well observable</td>
</tr>
<tr>
<td>2012-01-11T18:00Z</td>
<td>S.Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2012-01-13T18:00Z</td>
<td>South S.Scand</td>
<td>jet</td>
<td></td>
</tr>
<tr>
<td>2012-01-27T00:00Z</td>
<td>S.Scand</td>
<td>shear?</td>
<td>well observable, but long $\lambda_h$.</td>
</tr>
<tr>
<td>2012-01-29T12:00Z</td>
<td>Spitsbergen</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2013-01-03T18:00Z</td>
<td>S.Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2013-01-09T00:00Z</td>
<td>C.Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2013-01-12T06:00Z</td>
<td>Eureka</td>
<td>oro+jet</td>
<td>reasonably well observable</td>
</tr>
<tr>
<td>2013-01-20T00:00Z</td>
<td>C.Green</td>
<td>jet-exit ?</td>
<td>well observable</td>
</tr>
<tr>
<td>2013-01-20T18:00Z</td>
<td>W.Green</td>
<td>orography</td>
<td>mixture with synoptic structures</td>
</tr>
<tr>
<td>2013-01-29T18:00Z</td>
<td>South S.Scand</td>
<td>jet-exit ?</td>
<td>reasonably well observable</td>
</tr>
<tr>
<td>2014-01-09T00:00Z</td>
<td>Iceland</td>
<td>shear</td>
<td>well observable</td>
</tr>
<tr>
<td>2014-01-09T06:00Z</td>
<td>Iceland</td>
<td>shear</td>
<td>well observable</td>
</tr>
<tr>
<td>2014-01-24T18:00Z</td>
<td>S.Scand</td>
<td>orography</td>
<td>reasonably well observable</td>
</tr>
<tr>
<td>2014-01-25T00:00Z</td>
<td>S.Scand</td>
<td>oro+front</td>
<td>well observable</td>
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Table 4.4: List of prospective cases identified via the survey of 10 years of January by criterion maps: 2015

<table>
<thead>
<tr>
<th>GMT</th>
<th>location</th>
<th>likely source</th>
<th>comment</th>
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<tbody>
<tr>
<td>2015-01-01T06:00Z</td>
<td>S.,C.Scand</td>
<td>orography</td>
<td>well observable, short $\lambda_h$ in south, long north</td>
</tr>
<tr>
<td>2015-01-04T18:00Z</td>
<td>S.Scand</td>
<td>oro+front</td>
<td>well observable</td>
</tr>
<tr>
<td>2015-01-15T06:00Z</td>
<td>S.Green</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2015-01-20T06:00Z</td>
<td>S.Green</td>
<td>orography</td>
<td>well observable</td>
</tr>
<tr>
<td>2015-01-23T18:00Z</td>
<td>S.Green</td>
<td>oro+jet</td>
<td>well observable</td>
</tr>
<tr>
<td>2015-01-26T00:00Z</td>
<td>C.Scand</td>
<td>orography</td>
<td>well observable</td>
</tr>
</tbody>
</table>
5 Sensitivity of GLORIA towards Gravity Waves

The goal of this simulation study is to determine the capability of the GLORIA infrared limb imager to measure mesoscale GWs. The accuracy of reconstructing GW parameters, such as horizontal and vertical wavelength, amplitude and wave orientation is studied. For this purpose, an end-to-end simulation was performed which is described in the following. Figure 5.1 shows the concept of this end-to-end simulation.

\[
W_t^i = \hat{T} \cdot \sin (kx + \phi),
\]

where \(\hat{T}\) is the temperature amplitude, \(k \in \mathbb{R}^3\) the 3-D wave vector, and \(\phi\) the phase of the wave. In the present simulation study the temperature amplitude \(\hat{T}\) is arbitrarily chosen to be 3 K, the wave phase \(\phi\) to be 0°. For an improved conception, from now on, the 3-D wave vector \(k = (k_0, k_1, k_2)\) will mainly be expressed in terms of vertical
wavelength $\lambda_z = \frac{2\pi}{k_0}$, horizontal wavelength $\lambda_h = \frac{2\pi}{\sqrt{k_0^2 + k_1^2}}$, and horizontal wave direction $\varphi = \arctan2(k_1, k_0)$. Thus, if a downward pointing wave vector ($\lambda_z > 0$, upward propagating wave) is assumed, a wave with wave direction of 0° has east-west oriented wavefronts and is tilted towards the north, a wave with wave direction of 180° is tilted towards the south. These true waves are added to the climatological temperature field to gain the so-called true temperature field $T_t \in \mathbb{R}^m$.

From this predetermined atmospheric state and with a given flight path, the GLORIA measurement simulator calculates a set of infrared spectra, as would be measured by the GLORIA instrument (Krisch et al., 2018). A tomographic retrieval (Sec. 3.2) is performed using these simulated infrared spectra.

A background removal algorithm then subtracts the climatological temperature field $T_c$ from the retrieved temperature field $T_r \in \mathbb{R}^m$ to obtain the retrieved wave structure $w_r \in \mathbb{R}^m$. In a real measurement case the background field is unknown and has to be identified by mathematical filtering methods. To solely investigate the sensitivity of the measurement concept and exclude any additional effects, these filtering methods are not used for the simulation study.

Finally, the retrieved wave structure is compared to the original true wave structure. By repeating this process for different horizontal and vertical wavelengths, the observational filter of our measurement method is established. To interpret the retrieved wave structure with respect to gravity waves, the wave parameters amplitude, phase and wave vector have to be derived, using the small-volume few-wave decomposition method S3D (Sec. 3.6). Comparing these retrieved wave parameters to the prescribed true wave parameters gives detailed information on the usability of the different retrievals for GW research.

In this chapter, observational filters for various flight patterns are compared. Section 5.1 presents the sensitivity towards gravity waves for circular flight patterns, Sec. 5.2 for linear flight patterns.

### 5.1 Full angle tomography (FAT)

Former studies demonstrated the feasibility of 2-D tomography of rearward looking satellite instruments for the retrieval of 3-D atmospheric structures such as gravity waves (Ungermann et al., 2010a). However, the concept of 3-D tomography with sideways looking airborne instruments is quite different and, thus, may exhibit different characteristics. In 2-D tomography a volume is reconstructed from rearward looking measurements on a moving platform which slice the volume into multiple 2-D images. In 3-D FAT, a volume is reconstructed from measurements at different sides all around the volume. Thus, with FAT the problem of wave orientation with respect to instrument and flight path (Ungermann et al., 2010a) does not appear.
Figure 5.2: FAT retrieval and S3D results for a wave with 400 km horizontal, 6 km vertical wavelength, and horizontal wave direction $\varphi_t = 180^\circ$. Panels (a-c) show the true wave, panels (d-e) the retrieved wave, and panels (g-i) the difference between both. The black dashed lines mark the area covered by tangent points. In panels (j-m) the S3D results horizontal wavelength $\lambda_h$, vertical wavelength $\lambda_z$, horizontal wave direction $\varphi$, and temperature amplitude $\hat{T}$ are pictured.
Figure 5.3: Specific observational filters for (a) horizontal wavelength $\lambda_h$, (b) vertical wavelength $\lambda_z$, (c) temperature amplitude $\hat{T}$, and (d) horizontal wave direction $\phi$ for the FAT retrieval. The yellow lines mark errors of 10% for the horizontal wavelength (Panel (a)) and 20% for the amplitude (Panel (c)).
Figure 5.2 shows a 3-D FAT of a GW with horizontal wavelength of 400 km, vertical wavelength of 6 km and horizontal wave direction $\varphi = 180^\circ$. Pictured are three cross sections through the 3-D volume at 10.5 km altitude, at 0° N and at 0° E. The first column shows the true wave, the second column the retrieved wave and the third column the difference of both. Within the tangent point area (dotted lines) the temperature error is below 0.5 K. This is in good agreement with the determined accuracy of 0.5 K (Sec. 3.2).

A S3D fit (Sec. 3.6) was performed for this retrieval at 10.5 km altitude with a cube size of 5 km x 400 km x 400 km. The results of this fit can be seen in Fig. 5.2j-m. Within the hexagonal flight pattern the horizontal and vertical wavelength, and the horizontal wave direction are well reproduced. The original amplitude of 3 K is underestimated by 0.1 K.

These S3D results are used to construct the specific observational filters in Fig. 5.3. A mean value of the S3D fit results between 1° S, 1° N, 1° W, and 1° E gives the specific observational filter of the respective wavelength pair. The horizontal wavelength, the vertical wavelength and the horizontal wave direction are well reproduced for all tested waves. Further, there appears no phase shift in the FAT retrieval in contrast to conventional 1-D retrievals (Ungermann et al., 2010a). However, the amplitude of the waves is slightly reduced for waves with horizontal wavelength below 200 km or vertical wavelength below 3 km.

This simulation study shows, that FAT is able to properly reconstruct the wave vectors of mesoscale GWs. However, the observational filter of the temperature amplitude has to be taken into account, when comparing these measurements to different data sets.

5.2 Limited angle tomography (LAT)

5.2.1 Dependence of the retrieval results on horizontal and vertical wavelengths

Figure 5.4 shows a comparison of the LAT retrieval results for different wavelengths. The waves in column 1 and 4 have a larger horizontal wavelength of 600 km compared to the waves in columns 2 and 3 with 200 km horizontal wavelength. The vertical wavelength of 6 km of the waves in columns 1 and 3 is longer than the vertical wavelength of 2 km of the waves in columns 2 and 4. The waves with large vertical wavelength in columns 1 and 3 are well reproduced by the LAT retrieval within the tangent point covered area with errors below 0.5 K. The waves with short vertical wavelengths show larger temperature errors of up to 1.5 K within the tangent point area. This difference comes from the curved LOS through the straight wavefronts, which leads to an averaging over different wave phases. For the waves with short vertical wavelength the
LOS crosses multiple opposite wave phases, which decreases the measurement signal. A similar dependence of the sensitivity on the alignment of phasefronts with LOS was observed for sub-limb viewers (Wu and Waters, 1996; McLandress et al., 2000).

All retrieved waves show a slight V shape pattern, which is more emphasized for the waves with short vertical wavelength. This V shape is probably caused by the parabola shape of the LOS. The retrieval does not know, where along the line-of-sight how much of the measured radiation was emitted, unless crossing measurements give sufficient information. As the LAT has fewer measurements at different angles, the temperature signal is redistributed according to the weighting function (Fig. 3.2b) along the LOS. This can be nicely seen in the vertical cross sections in Fig. 5.4(g, o), where the warm temperature follows the LOS upwards behind the tangent point. This vertical shift of temperature also causes the northward oriented V shape in the horizontal cross sections.

As already for the FAT case, the specific observational filters were calculated using the S3D fits of the LAT retrievals (Fig. 5.5). The deviations of horizontal and vertical wavelengths are mainly below 10%. Only for very short vertical and very long horizontal wavelengths errors of above 20% appear. This is probably due to the above mentioned V shape deformation of the wave, which is more difficult to fit with one single sinusoidal wave. The same problem appears for the horizontal wave direction. For waves with short vertical and long horizontal wavelengths and, thus a strong V shape, the direction cannot be derived properly anymore. For the rest of the waves the direction error stays everywhere below 10°. The observational filter for the amplitude shows a similar pattern as for the FAT case.

**Dependence of retrieval results on the wave orientation**

Due to the limited measurement sector, the orientation of the wave with respect to the instrument position might be important for LAT. Figure 5.6 depicts the retrieval results for waves with horizontal wave directions turned by 30° ($\varphi_t = 210°$) compared to those in Fig. 5.4. In the vertical, the wavefronts are tilted southward and, thus, towards the instrument. They decrease in height with increasing distance to the flight path.

Overall, the structures are reproduced reasonably well. As for the perfectly perpendicularly-aligned waves already, waves with long vertical wavelengths (Fig. 5.6a–d and Fig. 5.6e–l) are reproduced better than waves with short vertical wavelengths (Fig. 5.6e–h and Fig. 5.6m–p).

Due to the tilt of the waves towards the aircraft, the LOS is partly aligned with the wavefronts before the tangent point. This effect is stronger for steep waves such as in Fig. 5.6k than for relatively flat waves such as in Fig. 5.7c, g and o. Due to this alignment the area of best sensitivity is shifted towards the aircraft for the steep wave. Spreading the signal now around this shifted sensitivity maximum, just spreads
Figure 5.4: Cross sections of retrieved waves (first and third row) and differences between true and retrieved waves (second and forth row) of the LAT retrieval. The different columns show waves with different horizontal and vertical wavelengths. The true horizontal wave orientation of all waves is $\phi_t = 180^\circ$ and, thus, these waves have wavefronts perpendicular to the flight path. The black dashed lines mark the area covered by tangent points. The grey line in the vertical cross sections indicates a LOS for a measurement with $90^\circ$ azimuth angle and tangent point altitude of 10.5 km.
Figure 5.5: Specific observational filters for (a) horizontal wavelength $\lambda_h$, (b) vertical wavelength $\lambda_z$, (c) temperature amplitude $\hat{T}$, and (d) horizontal wave direction $\varphi$ for the LAT retrieval and true horizontal wave orientation $\varphi_t = 180^\circ$ and, thus, these waves have wavefronts perpendicular to the flight path. The yellow lines mark errors of 10%, 10%, 20% and 10° in Panels (a-d), respectively.
the signal along the same wave phase, as the LOS has little curvature in this region. Therefore, no strong shape deviation is observed. For the flat waves a similar V shape can be observed as for the waves in Fig. 5.4 due to a spreading of signal along LOS around the tangent point.

In the observational filter (Fig. 5.7) a small decrease in the quality of amplitude reproduction can be seen compared to the observational filter of perfectly east-west aligned waves (Fig. 5.5). However, the wavelengths and wave direction are barely influenced and reproduced at a similar high quality. The V shape of the waves only occurs outside the tangent point region, thus proper horizontal wave directions can be observed.

Fig. 5.8 shows the retrieval results for waves turned by -210° ($\varphi_t = 30°$) compared to Fig. 5.4. These waves are tilted northward and, thus, away from the flight path. Only for the wave with large horizontal and large vertical wavelength (Fig. 5.8a–d) the temperature amplitude is reproduced well within the tangent point region. However, the horizontal orientation in this area, which should be similar to Fig. 5.6a from north-west to south-east is not recovered. The same happens for waves with short vertical wavelengths (Fig. 5.8e–h and m–p): The information about the horizontal wave direction is lost within the retrieval. Again a V shape appears for all these waves. Due to the inverse vertical tilt compared to Fig. 5.6 the opening of the V shape is this time to the south.

For steep waves (Fig. 5.8k) the main signal is again shifted, this time behind the tangent point area, where the LOS and the wave fronts are well aligned. Thus the spreading of the signal does not influence these waves as strongly as the flat waves and the horizontal orientation does not get lost in the retrieval. The decreased amplitude compared to Fig. 5.6 can be explained by the fact that the maximum of the weighting function along LOS is located slightly before the tangent point (Fig. 3.2a).

A similar picture is given from the observational filter in Fig. 5.9. Even though the amplitude is underestimated for very steep waves, the horizontal wave orientation can be derived accurately. However, the flatter the wave gets, the worse the derived horizontal wave direction. For waves with horizontal to vertical wavelengths ratio of above 200, the direction error exceeds 30°. Also the horizontal wavelengths reproduction is decreased somewhat compared to the two cases before (Fig. 5.5 and Fig. 5.7).

Further tests with horizontal wave direction $30° < \varphi_t < 90°$ and $210° < \varphi_t < 270°$ show a drastic decline in the amplitude sensitivity towards waves with short horizontal wavelengths. For waves tilted away from the flight path ($\varphi_t > 30°$) the fit quality of the horizontal wave direction and the horizontal wavelength decreases drastically already at $\varphi_t = 40°$.

These studies show that LAT applied to gravity waves gives best results for waves with wave fronts perpendicular to the flight path and, thus, horizontal wave vector $\varphi_t = 180°$. However, if the wave is slightly turned, the quality of the derived wave parameters is not affected strongly as long as the wave is tilted towards the instrument...
Figure 5.6: Cross sections of retrieved waves (first and third row) and differences between true and retrieved waves (second and forth row) of the LAT retrieval. The different columns show waves with different horizontal and vertical wavelengths. The true horizontal wave orientation of all waves is $\varphi_t = 210^\circ$ and, thus, these waves are tilted towards the aircraft.

(180° $\leq \varphi_t \leq 210^\circ$). In general, waves are best retrieved when their aspect ratio of horizontal and vertical wavelengths, i.e. their steepness, is favourable for an alignment with the LOS. In these cases, tilts towards and away from the instrument may give reasonable results.
Sensitivity of GLORIA towards Gravity Waves

Figure 5.7: Specific observational filters for (a) horizontal wavelength $\lambda_h$, (b) vertical wavelength $\lambda_z$, (c) temperature amplitude $\hat{T}$, and (d) horizontal wave direction $\varphi$ for the LAT retrieval with true horizontal wave orientation $\varphi_t = 210^\circ$ and, thus, waves, which are tilted towards the aircraft. The yellow lines mark errors of 10%, 10%, 20% and 10$^\circ$ in Panels (a-d), respectively.
Figure 5.8: Cross sections of retrieved waves (first and third row) and differences between true and retrieved waves (second and forth row) of the LAT retrieval. The different columns show waves with different horizontal and vertical wavelengths. The true horizontal wave orientation of all waves is $\varphi_t = 30^\circ$ and, thus, these waves are tilted away from aircraft.
Sensitivity of GLORIA towards Gravity Waves

Figure 5.9: Specific observational filters for (a) horizontal wavelength $\lambda_h$, (b) vertical wavelength $\lambda_z$, (c) temperature amplitude $\hat{T}$, (d) and horizontal wave direction $\varphi$ for the LAT retrieval with true horizontal wave orientation $\varphi_t = 30^\circ$ and, thus, waves, which are tilted away from aircraft. The yellow lines mark errors of 10%, 10%, 20% and 10$^\circ$ in Panels (a-d), respectively.
5.3 Conclusion

In this chapter we investigated the use of FAT and LAT applied to airborne limb imaging for gravity wave research. In contrast to FAT, which allows for the reconstruction of a large, cubic, 3-D volume, LAT can only reconstruct a band of 200 km around a banana-shaped vertical curtain parallel to the flight path. The horizontal resolution is 30 km in flight direction and 70 km perpendicular to flight direction. The vertical resolution is on the order of 400 m. This volume and resolution are sufficient to properly derive all important wave parameters such as the horizontal and vertical wavelengths, the amplitude, and the wave direction for waves with wavefronts perpendicular to the flight path. This is feasible due to the perfect alignment of wave phases and LOS and agrees well with earlier studies for other limb sounding concepts (Wu and Waters, 1996; McLandress et al., 2000; Unger mann et al., 2010a).

The quality of the 3-D reconstruction strongly depends on the orientation of the wave with respect to the instrument. If the waves are slightly turned away from the perfect orientation the quality of the derived wave parameters is not strongly affected as long as the wavefront is tilted towards the instrument. If the wavefronts are tilted away from the instrument, the retrieval will create artefacts which reduce the quality of the derived horizontal wave directions and wavelengths. For waves with horizontal wavelength under 300 km, the amplitude error is larger for waves with wavefronts tilted away from the instrument than for waves with wavefronts tilted towards the instrument. In general, the better the alignment of the wave phases and the LOS is, the more information is attained by the tomographic retrieval. Thus, steeper waves can be derived with better accuracy than flatter waves. For steep waves with a horizontal to vertical wavelength ratio below 200 correct wave directions can be derived independently of the tilt. However, for waves turned by more than 40° compared to the perfect, perpendicular case, the reconstruction quality decreases drastically for all tested waves.

In summary, for many GW cases the observation in LAT mode can be recommended. However, for short scale waves FAT is preferable due to the higher spatial resolution of 20 km x 20 km x 200 m. The slightly better accuracy of 0.5 K for FAT compared to 0.7 K for LAT also makes FAT favourable for low amplitude waves. Furthermore, when the precise orientation of the wave cannot be predicted before the flight, FAT should be the method of choice. Nevertheless, for many other cases, LAT might be preferred due to its shorter acquisition time.

The content of this chapter is published in [Krisch et al. (2018).]
6 Aircraft Campaign

6.1 Payload overview

The scientific payload of HALO encompasses three remote sensing instruments: GLORIA, the downward looking water vapor, cloud and ozone lidar WALES, and the differential optical absorption spectrometer miniDOAS. In addition, the BAHAMAS system as well as a number of in-situ instruments are part of the payload. The meteorological dataset (temperature, pressure and winds at high precision and high temporal resolution) from BAHAMAS as well as the in-situ trace species measurements are selected to characterize chemistry and exchange processes in the UTLS. The full list of instruments can be found in Table 6.1. In the following subsection we describe those instruments most relevant for gravity wave research.

6.1.1 GLORIA

GLORIA (Friedl-Vallon et al., 2014; Riese et al., 2014) is a cooled Fourier transform infrared (FTIR) spectrometer designed to fly on board the German research aircraft HALO or the Russian M55 Geophysica. It measures the infrared emission of atmospheric species in the spectral range from 780 cm\(^{-1}\) to 1400 cm\(^{-1}\). The radiation coming from the scene is modulated with a Michelson linear-slide Interferometer and imaged on a large MCT focal plane array. An integrated cooling system operating with solid carbon dioxide permits low temperature operation of the spectrometer and thus allows detecting the characteristic infrared emission spectral features of atmospheric species at good signal to noise ratio. The spectrometer is mounted in a three-axis gimbal to provide pointing stability and agility for different atmospheric observation modes. The instrument operates in an depressurised compartment and observes the atmospheric radiation through an opening on the side of the instrument bay. See Section 3.1 for details.

Intense simulations were carried out to demonstrate the feasibility to retrieve temperature structures of GWs from GLORIA radiances. The details of this work can be found in Chapter 5 of this report.
### Acronym Explanation

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Explanation</th>
</tr>
</thead>
<tbody>
<tr>
<td>AENEAS</td>
<td>NO&lt;sub&gt;x&lt;/sub&gt; measurement</td>
</tr>
<tr>
<td>AIMS</td>
<td>Atmospheric Chemical Ionization Mass Spectrometer</td>
</tr>
<tr>
<td>BAHAMAS</td>
<td>Basic Halo Measurement and Sensor System</td>
</tr>
<tr>
<td>Dropsonde System</td>
<td></td>
</tr>
<tr>
<td>FAIRO</td>
<td>Fast ozone measurement</td>
</tr>
<tr>
<td>FISH</td>
<td>Fast In-situ Stratospheric Hygrometer</td>
</tr>
<tr>
<td>GhOST</td>
<td>Gaschromatograph for Observation of Stratospheric Tracers</td>
</tr>
<tr>
<td>GLORIA</td>
<td>Gimballed Limb Observer for Radiance Imaging of the Atmosphere</td>
</tr>
<tr>
<td>HAGAR-V</td>
<td>High Altitude Gas AnalyzeR</td>
</tr>
<tr>
<td>HAI</td>
<td>Hygrometer for Atmospheric Investigation</td>
</tr>
<tr>
<td>TRIHOP</td>
<td>Three-channel tunable diode laser instrument for atmospheric research</td>
</tr>
<tr>
<td>miniDOAS</td>
<td>Differential Optical Absorption Spectroscopy</td>
</tr>
<tr>
<td>WALES</td>
<td>Water Vapour Lidar Experiment in Space</td>
</tr>
</tbody>
</table>

Table 6.1: List of scientific instruments for the mission.

<table>
<thead>
<tr>
<th>product</th>
<th>accuracy</th>
<th>precision</th>
</tr>
</thead>
<tbody>
<tr>
<td>temperature</td>
<td>0.5 K</td>
<td>&lt; 0.1 K</td>
</tr>
<tr>
<td>horizontal wind</td>
<td>1 ms&lt;sup&gt;-1&lt;/sup&gt;</td>
<td>0.2 ms&lt;sup&gt;-1&lt;/sup&gt;</td>
</tr>
<tr>
<td>vertical wind</td>
<td>0.5 ms&lt;sup&gt;-1&lt;/sup&gt;</td>
<td>0.1 ms&lt;sup&gt;-1&lt;/sup&gt;</td>
</tr>
<tr>
<td>pressure</td>
<td>&lt; 1 hPa</td>
<td></td>
</tr>
</tbody>
</table>

Table 6.2: Parameters and accuracy and precision of BAHAMAS data products.

#### 6.1.2 BAHAMAS

BAHAMAS is part of the extended standard instrumentation on HALO. HALO is equipped with a nose boom which allows for more accurate temperature and wind measurements than possible on standard airplane instrumentation. After post-flight processing, data are provided shortly after flight (1–2 days). The resolution of nominal data is ≈1 Hz, for turbulence studies experimentally higher temporal resolutions are recorded. The standard data, accuracy and precision are listed in Table 6.2.

#### 6.1.3 Dropsondes

Dropsondes are sensors on parachutes released from the aircraft, which send meteorological information to a receiver in the aircraft by telemetry. They register temperature, GPS position information from which horizontal winds are deduced, and pressure. A sonde takes 15 minutes from 14 km to the ground. A temporal sampling of 2 Hz corresponds to a vertical resolution of dz = 20 m.
6.1.4 WALES lidar

The airborne four-wavelength water vapor differential absorption lidar (DIAL) WALES was developed as an airborne demonstrator within the framework of the ESA Earth Explorer mission proposal “Water Vapor Lidar Experiment in Space” to investigate the feasibility of operating an active profiling DIAL system in space. The WALES DIAL system consists of two transmitters, both based on an injection-seeded optical parametric oscillator (OPO) pumped by the second-harmonic of a Q-switched, diode pumped Nd:YAG laser. Thus, WALES is capable of simultaneously emitting four wavelengths, three online and one offline, in the water vapor absorption band between 935 nm and 936 nm. The three online wavelengths achieve the necessary sensitivity needed for measurements over the whole range of tropospheric water vapor concentration. A complete water vapor profile of the troposphere is composed by using the information of the partly overlapping line contributions. The averaged pulse energy is 35 mJ with a repetition frequency rate of 2\times100 \text{Hz}. The vertical resolution of the raw data is 15 m.

In addition to the 935 nm channel, the receiver is equipped with polarization sensitive aerosol channels at 532 nm and 1064 nm, the first one with High Spectral Resolution capabilities using an iodine filter in the detection path (Esselborn et al., 2008). This allows for collocated measurements of humidity and optical depth, as well as studies of clouds and aerosol optical properties. For a detailed technical description see Wirth et al. (2009). In the POLSTRACC configuration, two of the water vapor channels, i.e.
those used for the lower troposphere, were replaced by ozone channels. In the transfer flight on 12 January 2016 the ozone channel failed and could not be repaired during PGGS\textsuperscript{1} campaign. For the GWEX flights therefore only the water and cloud polarization channel could be used.

### 6.1.5 Supporting Data: Falcon

The DLR Falcon has been used for several decades for research campaigns. The ceiling altitude of Falcon is FL420, but in polar research there is also a limitation by the coldest temperatures of -70°C in which the airplane can fly. Like HALO, Falcon possesses a nose boom. The Basis Instrumentation system implemented is MA6 (Messanlage 6), variables, accuracy and precision are the same as for the BAHAMAS system on HALO (cf. Table 6.2). As a remote sensing instrument, a 2µm Doppler wind lidar from the Institute of Atmospheric Physics at DLR samples the disturbed wind field and gravity waves. The wind profiles are retrieved from line-of-sight measurements of the Doppler lidar by the velocity-azimuth display (VAD) technique. The instrument performs a conical step-and-stare scan around the vertical axis with a nadir angle of 20° combined with the movement of the aircraft, this results in an acycloid scan pattern. The horizontal resolution of the wind profiles ($5$–$10$ km) is determined by the time needed for one scanner revolution (24 times 1 or 2 s as accumulation time per scan position, plus 6 s for the scanner motion) and the aircraft velocity ($160$–$240$ m s\(^{-1}\)). The vertical resolution of 100 m is determined by the pulse length of the laser (full width at half maximum (FWHM) = 400 ns ($\approx$120 m)) and the nadir angle of 20°. Nadir pointing measurements only without scanning mode will provide vertical winds in a horizontal resolution of $\approx$200 m.

### 6.1.6 Supporting Data: Ground-based measurements and soundings

In northern Scandinavia several observatories for the middle atmosphere exist. The sites are ALOMAR in Andenes, Norway, ESRANGE close to Kiruna, Sweden, and Sodankylä, Finland. From all sites weather balloons are launched regularly. ALOMAR and ESRANGE are equipped with radar and lidar systems for measuring winds and temperatures in the middle atmosphere. In Sodankylä, DLR deployed their mobile temperature lidar for the winter 2015/2016 (PGSG). Due to the unusually cold stratospheric temperatures in January 2016, the polar vortex was larger in extent. In addition, the phase of the planetary wave lead to a southward shift of the jet over Europe. As a result, there was no favourable condition for propagating GWs over northern Scandinavia and correlative measurements with the sites were not performed in January 2016.

\textsuperscript{1}PGGS stands for POLSTRACC/GW-LCycle/GWEX/Salsa.
6.2 Forecasting and flight planning

This section describes in detail the meteorological situation and following flight planning for the two GWEX flights on 25th and 28th of January 2016.

6.2.1 GWEX Flight 1 — 25 January 2016 (flight 10 overall)

Objectives

The target was to measure gravity waves generated by the interaction of low-level winds from a low-pressure system east of Greenland with the orography of Iceland. This follows Scenario 4 of the GWEX campaign implementation plan. The planned flight track is illustrated in Fig. 6.9.

Instrumentation

Primary: GLORIA, BAHAMAS, Dropsondes

Secondary: The full HALO payload for trace species measurements aboard HALO

GLORIA Instrument Operations

Dynamics mode was deployed in a hexagon north-east of Iceland supplemented by a linear flight cutting through the hexagon in a direction perpendicular to the main wave fronts. Drop sondes have been released to generate a curtain along this section.

Meteorologic Forecasts and GW situation

At low levels (Fig. 6.2c), a low-pressure system to the East of Greenland results in low-level winds at Iceland coming from the south. At higher altitudes (Figs 6.2a, b) the wind turns to a south-westerly. The wind directions are still sufficiently co-aligned to provide good propagation conditions for GWs. In particular no critical levels are expected.

The temperature residuals in Fig. 6.3 show wave signatures over and to the East of Iceland. For this region the GROGRAT projection (Fig. 6.4) indicates waves which can be measured by AIRS as well. The horizontal and vertical wavelengths (Figs 6.5 and 6.6) are sufficiently short to allow for a GLORIA hexagon flight. Also, the ground-based period is of the order of one day (Fig. 6.7), which means that the wave will not change largely during the two hours required for a hexagon. The tropopause is low (Fig. 6.8).

Due to the complex topography, complex waves are expected. This can be seen also from the temperature residuals in Fig. 6.3.
Figure 6.2: Geopotential height and wind fields from ECMWF at different levels in the stratosphere (a) and troposphere (b, c). Color code shows the absolute wind velocity, wind flags the direction and strength of the wind. Isolines indicate geopotential surfaces. The wind is generally directed in the direction of the isolines and strongest where the gradient is largest, i.e. in regions of dense isolines.
25–Jan–2016 12 GMT ; 12 km

Figure 6.3: Temperature residuals from ECMWF after subtracting zonal mean and global scale waves.

25–Jan–2016 12 GMT T_G= 1.0 T_A=

Figure 6.4: Criteria plot for GLORIA-measurable and AIRS measurable waves.
Figure 6.5: Horizontal wavelengths of GWs on 25 January 2016 12:00 UTC.

Figure 6.6: Vertical wavelengths of GWs on 25 January 2016 12:00 UTC.
25-Jan-2016 12 GMT; alt = 12 km

Figure 6.7: Ground based period of GWs on 25 January 2016 12:00 UTC.

25-Jan-2016 12 GMT

Figure 6.8: Thermal tropopause altitude on 25 January 2016 12:00 UTC.
Figure 6.9: Official flight plan for F10 as submitted to ATC.
6.2.2 GWEX Flight 2 — 28 January 2016 (flight 11 overall)

Objectives

The target was to measure GWs which are generated by the superposition of orography and spontaneous adjustment over southern Scandinavia. The planned flight track is illustrated in Fig. 6.17.

Instrumentation

Primary: GLORIA, BAHAMAS, WALES

Supporting: Falcon (wind lidar, MA6)

Secondary: The full HALO payload for trace species measurements aboard HALO

GLORIA Instrument Operations

Dynamics mode was deployed in the linear flight for long horizontal wavelengths. A Race-track pattern matches the low-level in-situ track with tangent points from the high-level leg. The flight was used to investigate two-sided linear-flight tomography.

Meteorologic Forecasts and GW situation

In the troposphere (Fig. 6.10c) we find two low pressure systems: An extended one at Iceland (65° N, 20° W) and a sharper one at 60° N, 30° E. This leads to a tropospheric jet with two maxima, one over the Atlantic Ocean south of Iceland and one over the Baltic States. In between, the jet is curved and divergent. Such a “jet-exit” region is a prominent source of GWs. On the late evening of 28 January, the jet-exit region is located over southern Scandinavia. At this time, both GW excitation by flow over orography and GW excitation by spontaneous adjustment in the jet-exit region exist in parallel. At both tropospheric levels, 700 hPa (Fig. 6.10c) and 500 hPa (Fig. 6.10b), the wind in this region is westerly, at 150 hPa (Fig. 6.10a) the wind is westerly with a weak southerly component. This provides favorable conditions for GWs to propagate to higher altitudes. Accordingly, the criteria map (Fig. 6.12) indicates GWs measurable by both GLORIA and AIRS.

The divergence of the isentropes is connected with a low tropopause altitude (see Fig. 6.16). This results in good measuring conditions for GLORIA because a low cloud top is expected. It also sharpens the tropopause. This could lead to partial reflection of GWs. In addition, the source and propagation conditions in the lower atmosphere are complex, because the weakest winds are found around 700 hPa.

The temperature residuals (Fig. 6.11) show south-north orientated wave fronts with a relatively long horizontal wavelength of approximately 400 km (cf. also Fig. 6.13).
The long horizontal wavelength cannot be well captured with a hexagon. Hence, we intend linear-flight tomography. By flying a race-track pattern (cf. flight plan in Fig. 6.17), tangent points from the southern leg match the in-situ and lidar observations on the northern leg.

The viewing volumes from the two legs are overlapping. For this two-sided, linear-flight tomography will be tested.

Due to instrument problems of GLORIA the flight was two hours delayed. This influenced one low-altitude and one medium-high flight leg. The two high-level legs, which are of particular interest for the GLORIA instrument, were performed as planned.
Figure 6.10: Geopotential height and wind fields from ECMWF at different levels in the stratosphere (a) and troposphere (b, c). Color code shows the absolute wind velocity, wind flags the direction and strength of the wind. Isolines indicate geopotential surfaces. The wind is generally directed in the direction of the isolines and strongest where the gradient is largest, i.e. in regions of dense isolines.
Figure 6.11: Temperature residuals from ECMWF after subtracting zonal mean and global scale waves.

Figure 6.12: Criteria plot for GLORIA-measurable and AIRS measurable waves.
Figure 6.13: Horizontal wavelengths of GWs on 28 January 2016 18:00 UTC.

Figure 6.14: Vertical wavelengths of GWs on 28 January 2016 18:00 UTC.
Figure 6.15: Ground based period of GWs on 28 January 2016 18:00 UTC.

28-Jan-2016 18 GMT ; alt= 12 km

Figure 6.16: Thermal tropopause altitude on 28 January 2016 18:00 UTC.
Figure 6.17: Official flight plan for F11 as submitted to ATC.
A mountain wave over Iceland propagates far away from its source

7.1 GLORIA observations

On the measurement day 25 January 2016, a southerly wind made landfall on the south coast of Iceland, thus exciting mountain waves. The wave structure over eastern Iceland was encircled by a hexagonal flight pattern with 460 km diameter between 10 and 12 UTC (Fig. 7.1). The aircraft flight altitude during this time was between 12.5 km and 13.5 km. Towards low altitudes, the GLORIA measurements were limited by clouds reaching up as far as 9 to 10.5 km. Before the hexagon a linear flight through the wave field has been performed to collect in-situ data at flight altitude and to release dropsondes.

For the GLORIA retrieval only the measurements taken between 10 and 12 UTC have been taken into account. The temperature residuals in Fig. 7.1 clearly reveal the complex 3-D structure of the wave field. An error analysis has been performed for this retrieval and the results are summarized in Table 7.1.

In Fig. 7.2, horizontal and vertical cross sections through the measurement volume are presented. They show how the wave structure varies with height and horizontal location. For instance, the wave fronts directly above Iceland (64°N to 65.5°N and 14°W to 18°W) are aligned east-west and tilted southwards against the prevailing southerly wind. Further to the north-east (65°N to 67°N and 10°W to 14°W), the horizontal orientation of the wave fronts turns more into south-west to north-east. The

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>accuracy</td>
<td>0.5 K</td>
</tr>
<tr>
<td>precision</td>
<td>0.05 K</td>
</tr>
<tr>
<td>vertical resolution</td>
<td>200 m</td>
</tr>
<tr>
<td>horizontal resolution</td>
<td>20 km</td>
</tr>
</tbody>
</table>

*Table 7.1: Results of the error analysis for the tomographic retrieval in Fig. 7.1 and Fig. 7.2. A detailed description how these parameters are calculated can be found in Section 3.2.*
horizontal wavelength varies inside the hexagon from 100 km up to 350 km. The vertical wavelength of the waves is between 3 km and 6 km. The temperature residuals range from ±4 K (in the south-west of the hexagon at an altitude of 12 km, 64°N to 65.5°N, and 14°W to 18°W) down to ±1 K (in the smaller scale waves in the north-western part of the hexagon at 66°N to 68°N and 16°W to 20°W).

In order to further interpret the GW structure and fully characterize it, wave parameters are derived using a small-volume few-wave decomposition technique (Sec. 3.6). Due to the prevailing wavelength range in our measurements (cf. Fig. 7.2), a cube size of 160 km x 160 km x 3.6 km, containing 4900 data points, was chosen. The fitted parameters are the 3-D wave vector $\mathbf{k} = (k, l, m)$ and the amplitudes (Fig. 7.3(c)). Horizontal and vertical wavelengths, the GWMF, and the horizontal wave direction were calculated from the wave vector $\mathbf{k}$ and shown in Fig. 7.3(a), 7.3(b), 7.3(d), and 7.3(e), respectively. Fig. 7.3(f) shows the orography of Iceland. The main mountain ridge is oriented in east-west direction. As expected for a mountain wave, the horizontal wave direction (Fig. 7.3(e)) is perpendicular to the ridge orientation.
A mountain wave over Iceland propagates far away from its source.

Figure 7.2: Horizontal (a-c) and vertical (d-f) cross sections through the 3-D volume shown in Fig. 7.1. The grey line marks the flightpath. The locations of the vertical cross sections are indicated by numbered dashed lines.
Figure 7.3: Three-dimensional sinusoidal wave fit of the GLORIA measurements in fitting cubes of 160 km x 160 km x 3.6 km at a center height of 11.5 km. Non-significant fitting results with wavelengths above 2.5 times the cube size are hashed. These parameters are used to drive the GROGRAT model, the results of which are shown in Fig. 7.4. Panel (e) shows the direction of the horizontal wave vector. Eastward direction corresponds to 90°, southward direction to 180°.
A mountain wave over Iceland propagates far away from its source

Integrating the GWMF over the horizontal extent of a GW event leads to the total momentum, which determines the maximal drag this GW event can exert on the background flow in coupling and dissipation processes. The total momentum of this GW case above Iceland is 3.1 GN. A comparison of all GW events in January 2016 has been performed in the 6-hourly operational analyses of ECMWF. The measured GW event can be classified as a very strong case since the sum of all occurrence probabilities of stronger events is well below 1%. This occurrence frequency is in good agreement with Alexander and Ortland (2010), Hertzog et al. (2012), and Podglajen et al. (2016) who present satellite and super-pressure balloon measurements at slightly higher altitudes.

In order to identify the GW source, we used the Gravity wave Regional Or Global RAy Tracer (GROGRAT; Sec. 3.7). Backward ray-tracing has been used in previous studies to locate GW sources (Preusse et al., 2014; Pramitha et al., 2015). In order to initialize a ray-tracer, the wave must be fully characterized. This capability is the main improvement of the GLORIA observations compared to previous remote sensing observations of temperature. GW parameters obtained from single vertical temperature profiles, lead to a cone of potential source regions instead of a precise source location (Gerrard et al., 2004). This is the reason why GWs derived from conventional limb scanner measurements have not been interpreted in terms of backward ray-tracing. Only the 3D nature and accuracy of the GLORIA measurements allows backtracing to the precise source location.

An error analysis revealed a systematic low bias of the vertical wavelengths due to the sinusoidal fitting routine (Krisch et al., 2017). Therefore, the vertical wavelengths from the sinusoidal fits were scaled by a factor 1.1, according to the determined bias, before being used for the ray-tracing.

Fig. 7.4 shows the backward ray-traces of the measured GWs from their measurement position (black crosses) down to the source location (red dots). The measurement position has been defined as the center point of the sinusoidal fitting cube. The strength of the GW is expressed by the size of the red dots, which has been chosen according to the GWMF at the source location. These GWMF values are conservative estimates, as the backward ray-tracing cannot account for dissipation processes. The source locations of the GWs, and in particular those of the highest GWMF, gather around the main mountain ridge of Iceland. The GWs are, thus, likely to have been excited by the southerly wind approaching these mountains. The ray-traces from the wave packet measured further in the north partly stop in the north of the island at single mountain peaks.

As can be seen in Fig. 7.4(c), the ray-traces need between 3 and 6 hours to reach the ground. This is in good agreement with a vertical group velocity of 2 to 3 km/h, which has been calculated from the measurements. Hence, the GWs are probably excited roughly 6 hours before the measurements were taken.

Forward ray tracing is used to examine the propagation of the GWs away from the measurement location (Fig. 7.4(b)). On the measurement day, the southerly wind
turned into a strong westerly direction above 10 km, creating a strong vertical wind shear. In this wind shear the GWs started to propagate eastward. This is confirmed by the measurements: at 11 km (Fig. 7.2(a)) the GWs are mainly located above the eastern part of Iceland, while at 13 km (Fig. 7.2(c)) the wave fronts already stretch far across the ocean. The waves require about one day to propagate to an altitude of 20 km (Fig. 7.4(c)). At the same time, they travel horizontally more than 2000 km (Fig. 7.4(b)). Over Eastern Europe, the GWs are refracted due to a horizontal wind shear and, thus, change their horizontal wave vector from southward to westward. This allows the waves to quickly propagate upward into the westerly wind in the mid stratosphere.

To mimic a typical GW parameterization scheme used in GCMs (McLandress, 1998), a second GROGRAT run (1D-GROGRAT) was performed with solely vertical propagation, time-independent background, and a horizontal wave direction constant with respect to altitude. In contrast to the full GROGRAT version (Fig. 7.4(c)), where the GWs propagate into the mid stratosphere, the GWs in the simplified version dissipate below 20 km (Fig. 7.4(d)). Two processes might play a significant role here: First, in the 1D GROGRAT version the GWs are not refracted and the wave vectors do not change its horizontal orientation with altitude. The westerly background winds at higher altitudes do not favor the propagation of GWs with wave vectors perpendicular to the wind direction. Second, in the full GROGRAT run, the GWs propagate horizontally away from the source. Hence, the GWs avoid the critical level positioned above the source location and more GWMF is transported to higher altitudes. Global mountain wave modeling (Xu et al., 2017) suggests that this effect may prevail also on a global basis.

Neither a realistic orientation of the wave vector, nor oblique GW propagation are incorporated in GW parameterizations used in current climate and weather prediction models (McLandress, 1998; Alexander and Dunkerton, 1999; Richter et al., 2010; McLandress et al., 2012; Garcia et al., 2017). However, both processes are context of several studies aiming to improve GW parameterizations (Preusse et al., 2009; Sato et al., 2009; Kalisch et al., 2014; Amemiya and Sato, 2016; Ribstein and Achatz, 2016; Garcia et al., 2017). The present paper provides a strong motivation to finally implement these processes in current climate and weather prediction models. Especially, as this could close gaps of GWMF in regions with sparse sources (McLandress et al., 2012) and reduce the cold-pole bias of climate and weather prediction models in the lower stratosphere (Garcia et al., 2017).

The content of this chapter is published in Krisch et al. (2017).
A mountain wave over Iceland propagates far away from its source

Figure 7.4: Ray traces calculated using the GROGRAT model. The starting positions of the rays are marked with black crosses and the grey line indicates the flight path. The size of the red circles in panel (a) indicates the GWMF at the end of the ray. Panel (a) shows the backward ray traces and panel (b) the forward ray traces, all starting at the measurement locations. Panels (c) and (d) show the change of GWMF with height for a full 4D GROGRAT (c) model run and a solely vertical 1D run (d).
7.2 Comparison to Satellite Instruments

The waves measured by the GLORIA instrument need more than a day to propagate to altitudes above 30 km. To compare the previous results to satellite measurements, we choose AIRS observations on 27 January 2016 during the descending orbit with an equator crossing at 01:30 LT. AIRS temperature residuals are shown in Fig. 7.5. Around 25°E, where we would expect to see some waves according to Fig. 7.4, temperature perturbations can be observed (Fig. 7.5). The sinusoidal fitting routine determines horizontal wavelengths of around 400 km and vertical wavelengths around 30 km (Fig. 7.6). According to the forward raytracing of the GLORIA measurements, waves with vertical wavelengths around 25 km and horizontal wavelengths between 250 km and 550 km are expected at the AIRS measurement altitude of 36 km. Also the amplitudes of the waves agree reasonably well with the forward raytracing. However, the huge distance between the source above Iceland and the measurement location at 36 km altitude as well as the long time period, which the waves need to propagate to this altitude, may introduce errors in the raytracing calculations (see Appendix of Krisch et al. (2017)). In addition, there are GWs above the British islands, which likely stem from local sources. This suggests that also the GWs above Scandinavia may in part stem from other sources. Thus, this comparison is by no means a prove that the waves observed with the AIRS instrument above Finland and the Baltic States are excited at the Icelandic mountains. However, it is a first indication and a promising sign for further studies.

Figure 7.5: Temperature residuals of the AIRS retrieval for the descending orbits with equator crossing time at 01:30 local time on 27 January 2016.
A mountain wave over Iceland propagates far away from its source.

Figure 7.6: Three-dimensional sinusoidal wave fit of the AIRS measurements in fitting cubes of 300 km x 250 km x 20 km at a center height of 36 km.
8 Flight 11 above southern Scandinavia on 28 January 2016

8.1 GLORIA observations

On the measurement day 28 January 2016, a westerly wind provided favorable conditions for mountain wave excitation at the Scandinavian Mountains. The wave structure over southern Scandinavia was probed with a race track pattern at different altitudes. GLORIA was operated during two linear flight legs with 700 km length each, which were oriented perpendicular to the predicted wave front orientation. These two legs are roughly 200 km apart and were performed in a clockwise orientation. The southern leg was performed between 20:00 UTC and 21:00 UTC from east to west. The northern leg was flown between 21:00 UTC and 22:00 UTC from west to east.

Figure 8.1 shows in the left column the temperature residuals derived from tomographic retrievals using measurements performed during the southern flight leg and in the right column temperature residuals derived from measurements taken on the northern flight leg. An error analysis has been performed for these retrievals and the results are summarized in Table 8.1. Both retrievals show a prominent wave structure with around 400 km horizontal and around 7 km vertical wavelength. This large scale structure is perturbed by a smaller scale wave with longer vertical but shorter horizontal wavelength. Even though the main characteristics are similar for the two retrievals, some differences between the two legs are apparent. The wave seems to change its horizontal orientation over the duration of the flight. This difference in direction is also indicated by the result of the sinusoidal fit results presented in Fig. 8.2, where a change of the wave orientation from $\varphi = 270^\circ$ to $\varphi = 290^\circ$ can be observed. Also the wavelengths vary. For the southern leg, the waves decrease in steepness from west to east, which can also be seen in the vertical cross section of the temperature residuals (Fig. 8.1(e)): At 200 km the waves have shorter horizontal and longer vertical wavelengths than at 600 km. For the northern flight leg, the waves are in general less steep than the waves observed during the southern flight leg.
<table>
<thead>
<tr>
<th>Quantity</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>accuracy</td>
<td>0.7 K</td>
</tr>
<tr>
<td>precision</td>
<td>0.05 K</td>
</tr>
<tr>
<td>vertical resolution</td>
<td>500 m</td>
</tr>
<tr>
<td>horizontal resolution across flight track</td>
<td>75 km</td>
</tr>
<tr>
<td>horizontal resolution along flight track</td>
<td>30 km</td>
</tr>
</tbody>
</table>

Table 8.1: Results of the error analysis for the tomographic retrieval in Fig. 8.1. A detailed description how these parameters are calculated can be found in Section 3.2.

Figure 8.1: Temperature residuals of the GLORIA tomographic retrieval for the flight on 28 January 2016 over southern Scandinavia. Shown are horizontal (a-c) and vertical (d-f) cross sections. The vertical cross sections (d-f) are along the dashed lines in (a-c). The gray line indicates the flight path.
Figure 8.2: Three-dimensional sinusoidal wave fit of the GLORIA measurements at a center height of 11.4 km in fitting cubes of 400 km x 250 km x 4 km with a tangent point weighting according to Sec. 3.6. In order to get accurate amplitudes, an amplitude and phase refit has been performed in fitting cubes of 100 km x 250 km x 1 km. The fitting results are used to drive the GROGRAT model, the results of which are shown in Fig. 8.3. Panel (b) shows the direction of the horizontal wave vector. Eastward direction corresponds to 90° and southward direction to 180°.
To identify the sources of these waves, the GROGRAT ray tracing model has been initiated with the derived wave parameters from Fig. 8.2. The results are shown in Fig. 8.3. The red dots indicate the end of the ray at the surface. Ray endings which do not reach the surface are marked with an empty circle. The size of the red dot or empty circle indicates the GWMF at the measurement location (thus at 11.4 km): the larger the circle the higher the momentum flux. Most of the rays and especially those with high GWMF values are traced back to the Scandinavian Mountains. However, some rays and especially those not reaching the surface indicate wave propagating from origins over the North Sea, where the back traces spread over a larger area. According to the meteorological forecasts (Sec. 6.2.2) a jet-exit region was moving over this area during the 28 January 2016, which might be a second source for the observed GWs. Thus, the main source is the orography, but also spontaneous adjustment in the jet exit region seems to play a role.

Forward ray tracing shows, that the waves propagate somewhat to the north and up to high altitudes (Fig. 8.3(b)). The GWMF is conserved up to an altitude of roughly 40 km, where the stratopause is located (Fig. 8.3(f)). There many rays deposit a major part of the GW momentum flux they carry.

As mentioned before, the temperature residuals measured by GLORIA are perturbed by smaller scale waves. To identify these smaller scale waves, the identified large scale waves have been subtracted from the temperature residuals in Fig. 8.1, which leads to the remaining temperature residuals in Fig. 8.4. Here, waves with amplitudes up to 1.5 K with short horizontal and long vertical wavelengths can be seen. Sinusoidal fits of these temperature fields are shown in Fig. 8.5. For the southern leg there are two wave packets identified. The first one is located around 60.5°N and 7°E and has horizontal wavelengths around 100 km and long vertical wavelengths. The second wave packet can be found around 60.5°N and 13°E. This wave packet has longer horizontal wavelengths on the order of 300 km. Both wave packets have a horizontal direction of 270°, i.e. due west. For the northern leg, only the wave packet around 60.5°N and 13°E can be separated from the background. However, the wave orientation is now around 90°. However, waves propagating towards the east would need to propagate much faster than the prevailing eastward wind and thus have very high phase speeds in a ground-based reference frame. Therefore a wave propagation direction of 270° is far more likely. The most likely explanation for the direction reversal are fit uncertainties due to the almost vertical wave fronts. In this condition the sinusoidal fit might have difficulties to identify the correct tilt of the waves. An alternative explanation would be a downward propagating wave (e.g. from wave reflection), with upward pointing k vector. The in depth analysis of these smaller scale waves, their sources, and propagation is still ongoing.
Figure 8.3: Ray traces calculated using the GROGRAT model. Panel (a) shows the backward ray traces and panel (b) the forward ray traces, all starting at the measurement locations. Panels (c-h) show the change of wave parameters and atmospheric stability with height.
Figure 8.4: Remaining temperature residuals of the GLORIA tomographic retrieval after subtraction of the wave of Fig. [8.2]. Shown are horizontal (a-d) and vertical (e, f) cross sections. The vertical cross sections (e, f) are along the dashed lines in (a-d). The gray line indicates the flight path.
Figure 8.5: Three-dimensional sinusoidal wave fit of the remaining GLORIA residuals shown in Fig 8.4 in fitting cubes of 100 km x 100 km x 1 km at a center height of 11.4 km.
8.2 Comparison to Satellite Instruments

The waves measured by GLORIA propagate up to altitudes of 40 km and more. Thus, a comparison with satellite measurements seems feasible. According to Fig. 8.3, the waves take between 3 and 12 hours to propagate up to 36 km. Thus, the AIRS measurements of the descending orbit on 29 January were chosen for the comparison. Figure 8.6 shows the temperature residuals measured by the AIRS instrument on 29 January 01:30 local time. Over Scandinavia this is around 03:00 UTC, which is 6 hours after the GLORIA measurements were taken. The forward ray tracing predicts GWs with amplitudes around 10 K above middle and northern Scandinavia (Fig. 8.3). However, the AIRS measurements barely show temperature fluctuations in this region. This is also seen in the S3D fits given in Fig. 8.7. These show very low amplitudes but quite long vertical wavelengths over northern Scandinavia. According to the forward ray tracing the waves are expected to have vertical wavelengths in the range of 5 km to 15 km. This seems in contradiction to the AIRS results. However, if the shorter vertical wavelengths were true, then the GWs may be hardly observable by the AIRS instrument due to the broad vertical averaging kernel of AIRS, which has a width of ~10 km (cf. Chapter 3, Section 3.3). As the wave can be observed in the ECMWF model at GLORIA flight altitude, we use ECMWF forecast date to study the AIRS visibility limits in more detail.

Figure 8.6: Temperature residuals of the AIRS retrieval for the descending orbits with equator crossing time at 01:30 local time on 29 January 2016.

Figure 8.8 shows the ECMWF forecast for 3 UTC on 29 January 2016 (left column).
At 36 km altitude a large wave field with amplitudes above 6 K can be seen above all over Scandinavia. Applying the AIRS averaging kernel to this model data yields in the temperature structure which the AIRS instrument would observe. This field is shown in the right column of Fig. 8.8. Above middle and northern Scandinavia the wave signatures vanish, above the south tip of Sweden some waves remain with strongly decreased amplitudes. This simulated observation of residual temperatures agrees very well with the real measurements (Fig. 8.6).

The sinusoidal fits of these temperature fields (Fig. 8.9) do not only confirm this picture but help explaining the situation. The vertical wavelengths of the ECMWF field (Fig. 8.9(c)) are below 15 km almost all over Scandinavia. Just above the very southern tip of Sweden the vertical wavelength raises to above 20 km. This area nicely co-locates with the area of increased amplitude in the AIRS observations (Fig. 8.9(e)). Interestingly, the AIRS observational filter does not only influence the observed amplitude but also the vertical wavelength. This can be seen in Fig. 8.9(c) and (g). The sinusoidal fits of the simulated AIRS observations (Fig. 8.9 right column) yield much longer vertical wavelengths in agreement with the real measurements (Fig. 8.7).

In conclusion, the large GW field above Scandinavia can barely be observed by the AIRS instrument due to its coarse vertical resolution. In contrast, a space born IRLI has a fine vertical resolution and could fully resolve these waves. For instance, the PREMIE concept had a point spread function of 1 km fully capable of observing this GW event.
Figure 8.8: ECMWF forecast initialized on 29 January 2016 midnight for 03 UTC and the influence of the AIRS observational filter. The left column shows the original ECMWF forecast data, the right column shows what remains if the model data is multiplied with the AIRS averaging kernel matrix.
Figure 8.9: Sinusoidal fits of the temperature residuals shown in Fig. 8.8 in fitting cubes of 300 km x 250 km x 20 km at a center height of 36 km.
9 An Attempt to Test the Polarization Relations

9.1 Introduction

What is required for studying global scale dynamics is the flux of GW pseudo momentum $F_p$ \cite{Fritts2003}. Of course we want to confine $F_p$ by observations. In principle, one can infer momentum flux directly, if one measures all three wind components. For the real world this proves to be difficult. We have encountered in this report already the concept of the observational filter and the necessity to remove the background, which adds to this observational filter and introduces errors. Momentum flux is defined as the average over one period or wavelength, so you have to use knowledge on the wave field in order to find a suited averaging length as well as a well-defined section through the wave field such as a truly vertical or horizontal profile. The latter is far from trivial as for instance a radiosonde drifts with the wind and an airplane (as we will see below) varies its altitude. Finally, pseudo momentum flux is related to momentum flux from winds via a factor $1 - \frac{f^2}{\hat{\omega}^2}$ which can be determined only via detailed wave analysis.

A more practical problem is that only very few techniques provide all three wind components with sufficient accuracy. Therefore the evaluation of most observational techniques relies on the polarization relations in some way or other. For radiosondes, dropsondes and lidar data, for instance, the hodograph technique (based on polarization relations) is used to determine the intrinsic frequency of the waves and the horizontal propagation direction. Having these, one can use the dispersion relation to calculate the horizontal wavelength. Finally, the phase relation between winds and temperature (again assuming polarization relations) is used to remove an 180° ambiguity in the horizontal propagation direction.

Polarization and dispersion relations rely on WKB linear theory. Are they still valid for large amplitude waves in a shear environment?

This has been studied by theoreticians. However, given that this is an essential question right on the basis of the majority of gravity wave analyses, very few experimental studies exit. Among the few examples are the following studies. Using the dispersion relation \cite{Eckermann2001} and \cite{Preusse2002} calculated the expected vertical wavelengths of mountain waves and compared them to CRISTA observations.
Stober et al. (2013) calculated expected horizontal wavelengths from hodograph analyses and compared them to the observed wavelengths in 3-D radar data. (Geller and Gong, 2010) investigated GWMF from radiosonde data using different techniques and different observables. They find that different parts of the wave spectrum are manifested, depending on the variables used. Calling to one’s mind the polarization relations this is not so surprising: the horizontal wind amplitude and the temperature amplitude can be related under typical stratospheric conditions by $\hat{u}|| \approx 2\hat{T}$ ($\hat{u}||$ in units of m/s and $\hat{T}$ in units of K). However, the vertical wind amplitude is related via $\hat{w} = \frac{\nu}{\kappa} \hat{u}||$. Thus, short horizontal wavelengths with an aspect ratio of $\lambda_z/\lambda_h \approx 1$ tend to be very prominent in the vertical wind, while the mesoscale waves with aspect ratios of $\lambda_z/\lambda_h < 0.1$ are more visible in temperature and horizontal winds.

In principle, the GWEX data provide us a unique opportunity to check on the polarization relations. We have 3-D observations from GLORIA and we have a complete set of variables ($u$, $v$, $w$, $T$) from the in situ observations. However, none of the observations is complete, in that it provides 3-D data and a full set of observables. Can they be utilized together to confirm polarization relations?

A further approach might be via trace species. Wüst and Bittner (2006) claim that GWs can be extracted from trace species perturbations, but with very little tests to their basic assumption: the trace species perturbations introduced via the vertical displacement need to be larger than the structures by horizontal transport. This is valid for shorter horizontal wavelength GWs in the mesopause OH layer (e.g. Eckermann et al., 1998), but is it valid also in the UTLS?

In this way, the following section will leave our existing knowledge and is an highly experimental attempt to merge very different observation techniques and observables.

9.2 Consistency of Variables in ECMWF Model Data

In a first step we test our approach on ECMWF data. The general circulation model (GCM) of the ECMWF integrated forecasting system (ECMWF-IFS) is a high resolution global model capable to resolve a larger part of the GW spectrum. The model is based on first principles and does not invoke any explicit GW physics. The ECMWF data have the following advantages: They

1. provide all observables,
2. in 3-D,
3. on a regular grid for a given specified time,
4. and with a common filter given by the spatial resolution of the model.

This means, respectively, that we can...
1. check the polarization relations,
2. choose the way how to evaluate the data,
3. do not have to worry about sampling issues,
4. and see only the mesoscale part of the spectrum.

A limitation on the part of the spectrum resolved seems a disadvantage rather than an advantage at first, but we will see in the following sections that it is beneficial for the data evaluation.

9.2.1 Background Removal

The first step of data evaluation is the background removal. We use global background removal of zonal waves smoothed in latitudinal direction with a 3rd order polynomial over a range of 5° latitude. In general the need to take into account higher wavenumbers increases at lower altitudes. In the mid stratosphere a zonal wavenumber of 6 is sufficient to isolate GW perturbations from the background. Testing the background removal for altitudes between 9 and 15 km for different zonal wavenumbers, we find that for temperature and vertical wind the background removal converges for a zonal wavenumber of 18, but that for the horizontal winds a wavenumber of 36 is required to isolate the GW induced mesoscale structures. This is shown in Fig. 9.1 which compares for 10 km, 12 km and 15 km altitude temperature and meridional wind residuals after background removal with zonal waves up to 18 and 36, respectively. There is little difference between the temperature residuals at one altitude but considerable difference for the winds. Only with a wave-36 detrending the winds exhibit similar patterns than the structures in the temperatures. Comparing the different variables (not shown) we find that above the tropopause vertical winds are most robust, temperatures are well detrended by a wavenumber 18, zonal winds require a wave-36 background removal and meridional winds are most demanding.

9.2.2 Application of S3D

For the investigations in this chapter, we use a two-step approach for the S3D. In the first step we perform an unweighted fit and determine the wave vector, in the second step we employ a Gaussian weight and are fitting amplitude and phase only. This provides reliable estimates for the wave vector and allows to attribute the local value of a varying amplitude field to the cube center. For each S3D fit we posses temperature amplitude \( \hat{T} \) and 3-D wave vector. From these we can calculate the wind amplitudes \( (\hat{u}_\parallel, \hat{u}_\perp, \hat{w}) \) of the GW. Using the phase information the values at the cube center are reconstructed and horizontal winds are projected to zonal and meridional direction.
Figure 9.1: Comparison of background removal for temperature (the two columns on the left) and meridional wind (the two columns on the right) based on a removal of zonal waves up to wavenumber 18 (respective left) and wavenumber 36 (respective right). Different rows show altitudes of 10 km, 12 km and 15 km.
9.2.3 GW Perturbations

Iceland, 25 January 2016

The polarization relations are introduced in chapter 2, section 2.2. According to the polarization relations one can calculate the perturbations induced by a GW for all observables (i.e. $T, u, v, w$), given that one possesses one of these variables and has a full characterization of the wave e.g. in terms of the 3-D wave vector. Since the background of this study is tomography of atmospheric temperatures, we start also for the model data from 3-D fields of temperature residuals. The temperature residuals are analyzed by the S3D tool using overlapping cubes. For this part of the study, we perform S3D fits for each grid point of the residual data - minus the edges.

Figure 9.2 shows the results from this analysis and compares residuals directly from the ECMWF (i.e. without invoking any GW physics) with the reconstructed perturbations from the polarization relations. The first row compares the original temperature residuals (panel a) with reconstructions from the S3D based on one wave component (panel b) and two superposed waves (panel c). The comparison shows that already the first wave component captures the major features and that the agreement further improves by using two wave components. The three rows below show zonal, meridional and vertical wind, respectively. As was shown in Chapter 7, the wave vector is directed towards south-south-east. For this wave event we expect the wind component parallel to the wave vector $u_\parallel$ to be much larger than the wind component perpendicular $u_\perp$. The direction of the wave vector and the dominance of $u_\parallel$ lead to a much stronger pattern of the wave in the meridional wind than in the zonal wind. We find this both in the direct residuals (left column) as well as in the reconstructed GW perturbations (middle and right column). Overall, the agreement of the reconstructed wave pattern is good, both in amplitude as well as in the shape of the main wave pattern. Already the first wave component captures most of the variations. While zonal and meridional wind tend to emphasize the same wave patterns as the temperature residuals, vertical winds tend to emphasize shorter horizontal scale GWs. Also for the vertical wind the salient features are reproduced. However, there is a much larger contribution of the second wave component (cf. panels k and l) and in general the amplitude in the main wave packet remains underestimated. On the other hand, there is some spread of the large scale waves into regions where there is almost no signal in the direct residuals in panel j. This is likely due to the finite size of the S3D analysis cubes.

Overall, we find consistency between the different observables via the polarization relations. All major structures are reproduced and the size of the variation is in good

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1In order to fill a cube symmetrically around the cube center, this cube center has a minimum distance of half a cube size from the edge of the volume for which residuals are provided. For ECMWF that restricts mainly altitude. Avoiding to include the tropopause into the analysis volume means that the lowest cube center is located half a vertical cube size above the tropopause.
**Figure 9.2:** Comparison between direct ECMWF residuals (left row) and reconstructions from S3D fits of temperatures for the first wave component (middle row) and the superposition of first and second wave component (right row). Panels b and c show direct reconstructions from the S3D fits, all lower rows give reconstruction of wind perturbations based on the temperature fits and the polarization relation. The good agreement of the reconstructed wind perturbations with the ECMWF residuals confirms the polarization relations. Grey stippled areas indicate regions where the fits are less reliable and the resulting vertical wavelength exceeds 2.5 times the cube size.
An Attempt to Test the Polarization Relations

agreement. It should be noted that this is not a trivial result since ECMWF fields are not based explicitly on any GW physics.

Scandinavia, 28 January 2016

For the second flight on 28 January 2016 above Southern Scandinavia GW fronts are oriented North-South parallel to the mountain ridge and opposite to the background wind. This situation can be well captured by the vertical cross section along the $60^\circ$ latitude circle shown in Fig. 9.3. In this case (cf. Chapter 8) there is a superposition of large scale waves with horizontal wavelengths of $\sim 400$ km and much shorter scales. This is difficult to capture with a single cube size and we here focus on the first wave component. As for Fig. 9.2 the left column of Fig. 9.3 shows the direct residuals from ECMWF after background removal. The middle column shows the reconstruction from the first wave component. For the temperatures shown in Fig. 9.3e wave vector $k$, amplitude $\hat{T}$ and phase $\phi_T$ are adapted in the fit. For the winds, the wave vector is taken from the temperature fit, but amplitudes $\hat{u}$, $\hat{v}$ and $\hat{w}$ as well as phases $\phi_u$, $\phi_v$ and $\phi_w$ are fitted. The right column (panels i to k) shows the reconstructed GW perturbations from the polarization relations. These are also shown in the middle column as black contour lines, solid for positive, dashed for negative.

The observed large scale wave is propagating almost due west. Therefore the wind component along the wave vector $u_||$ results in the zonal wind perturbation, the Coriolis-force-induced wind component $u_\perp$ results in the meridional wind perturbation. The zonal wind residuals (panel b) are well reproduced by the fits (panel f) and the GW perturbations from the polarization relations (panel i). Comparing the color shading and the contour lines in panel f, there is a minor phase shift between the expected and fitted wind structures. Filtering only the meridional wind components which match the wavelengths found in the temperatures, there is reasonable agreement between the fits (panel g) and the reconstructed GW perturbations from the temperatures (panel j). However, it is evident that the residuals contain some larger scale structures not visible in the GW patterns. This points likely to some non-GW phenomenon. We will come back to this when discussing the trace gas signatures. For vertical winds the fitted residuals (panel h) and the GW perturbations from the polarization relation (panel k) are in good agreement. They also well reproduce the large scale features of the residuals given in panel d. In addition, vertical winds highlight smaller horizontal scale GWs which are much less pronounced in temperature and horizontal winds.

The GW perturbations deduced from temperature via S3D fit and the polarization relations match zonal wind residuals very well and explain the large scale patterns in the vertical wind velocities. Vertical wind velocities show signatures from smaller scale waves in addition. Meridional winds can be explained partly by the large scale GWs but show also features which likely are not due to GWs.
Figure 9.3: Vertical cross sections along 60° latitude of temperature perturbations (upper row), zonal wind perturbations (2nd row), meridional wind perturbations (3rd row) and vertical wind perturbations (bottom row). Left column shows direct residuals after background removal, center column shows fits with wave vectors $k$ from fits of temperature residuals, and right column GW perturbations calculated via the polarization relations. In the center column contour lines repeat the GW perturbations from the right column visualizing the wave phase.
9.3 Comparison to aircraft in situ data

9.3.1 Comparison to aircraft in situ data: Iceland case

In order to interpret the in situ observations taken by BAHAMAS, the HALO meteorological basis sensors, straight flight segments need to be considered. Therefore the whole flight path is analyzed and divided into straight flight segments. The flight segments relevant for the Iceland case are shown in Fig. 9.4. The first segment of relevance for the considered GW is segment 6, which crosses the hexagon from north-west to south-east, roughly in opposite direction to the background wind and perpendicular to the wave fronts. Segments 8 to 13 form the sides of the hexagon. Figure 9.4b shows the altitude profile along the flight. The autopilot of the plane intends to keep HALO on constant pressure. In a background pressure gradient this leads to changes in geometric altitude. For instance, HALO climbs about 250m during segment 6 and descends again until end of segment 10. There HALO ascends to a higher flight level, which is maintained until the end of segment 13. Shorter scale variations are due to small-scale waves and turbulence.

![Figure 9.4: Map of Iceland with the flight path of HALO and labeled straight flight segments (panel a). In panel b the altitude profile of the flight path is shown.](image)

The atmospheric temperature versus flight time is shown in Fig. 9.5. The black line indicating the smooth background shows considerable variation due to a) variations in altitude and b) variations in geographical location. Superposed on this are wave perturbations from GWs on different scales. In order to compare these variations we subtract the same smooth background from ECMWF from all these data sets. The advantage of this approach is that we do not introduce different structures into these data by subtracting different background estimates. In other words, if the background
itself shows signatures of waves, this influences all data sets in the same way. The disadvantage of this method is that the individual observations may have biases with regard to ECMWF. With a single flight we cannot apply statistical corrections. Such biases therefore will be visible in difference plots.

Figure 9.6 shows the variations of temperature (panel a), zonal wind (panel b), meridional wind (panel c) and vertical wind (panel d) after subtraction of a smooth background from ECMWF. The data are plotted versus the total flight distance after take off. For the BAHAMAS data also a low-pass filter was applied (dark green curve); the response function is given by the blue curve in Fig. 9.7. The filter removes waves shorter than 100 km wavelengths, which are not visible to GLORIA. The variations observed by GLORIA (orange) and smoothed BAHAMAS (dark green) data are of similar size. Particularly large variations are, for instance, along segment 6 and the peak at the start of segment 10. ECMWF residuals (light blue) are somewhat smaller in amplitude but generally in good agreement. The reconstructed perturbations from S3D (red for GLORIA and dark blue for ECMWF) smooth these variations. For GLORIA this effect is emphasized since for segments 8 to 13 (the hexagon edges) the fit volumes are only partially covered by tangent points. In segment 6 there is a phase shift between GLORIA and BAHAMAS with some delay in GLORIA. Also there are slight differences in shape: GLORIA and ECMWF show a stronger minimum around 2000 km flight distance, but a less pronounced one around 2200 km flight distance.

As discussed in Section 9.2.3 we expect to see the GW perturbations most clearly in the meridional wind (panel c). Indeed we find that the zonal winds reconstructed from GLORIA (red) and ECMWF temperature fits (dark blue) resemble the BAHAMAS observations reasonably well. Differences are consistent to those in temperature, such as the phase delay in the GLORIA data and the deeper first minimum in GLORIA and ECMWF compared to BAHAMAS. Also the peak at the start of segment 10 is in good agreement. Zonal winds are less clear: the peak at the start of segment 10 is again in good agreement, but there are differences for segment 6. Actually, for ECMWF direct residuals and S3D reconstruction also do not match very well. This indicates the presence of processes not well described by the GW polarization relations. The vertical wind fluctuations (panel d) are dominated by very short scales. In order to keep the vertical wind variations due to the longer scales visible, the y-axis is limited to ±1 ms⁻¹. The shorter scales exceed this. A few features, such as the peak at the start of segment 10 are in agreement between BAHAMAS and the other data sets. However, for most patterns, there is very little similarity.

The most likely reason for this discrepancy is the dominance of small scales in the vertical wind fluctuations. We try to mitigate this by smoothing the BAHAMAS data. However, we can perform this only along the flight track. The larger scale waves see the mountains at the southern coast as a ridge. However, the topography of Iceland consists of volcanoes and is far more complex. Due to this complex topography, shorter scale waves can have any propagation direction which still maintains some southward
Figure 9.5: Temperature versus flight time for the GW part of the Iceland flight. Different colors indicate different data sets. Shown are BAHAMAS data (light green), ECMWF data (light blue), ECMWF data reconstructed from the S3D fit (dark blue), GLORIA data (orange) and GLORIA data reconstructed from S3D (red). The black line indicates the smooth background generated from ECMWF data by the low-pass filter retaining only global waves with zonal wavenumbers 0-18.
Projected to the flight track the wavelength seen by BAHAMAS will, in general, be substantially longer than the true wavelength. The filter is hence of limited use.

Figure 9.6: Comparison of temperature and winds from various data sets, Iceland, 25 January 2016. Shown are perturbations from BAHAMAS (light green), smoothed BAHAMAS data (dark green), ECMWF (light blue), S3D reconstruction from ECMWF (dark blue), GLORIA (orange) and S3D reconstruction from GLORIA (red). Panel a shows perturbations for temperature, panel b for zonal wind, panel c for meridional wind and panel d for vertical wind. For discussion see text.

In conclusion, GLORIA, BAHAMAS and ECMWF temperatures show the same salient features. The main wave patterns in segment 6 and for the start of segment 10 agree well. The zonal winds are consistent between observations and reconstruction from the S3D fit via the polarization relation. This confirms that the analyzed temperature structures are GWs. It also confirms the magnitude of the amplitudes. This is also a somewhat indirect validation for the size of the momentum flux. For a direct validation we would need consistent vertical winds. These, however, are dominated by shorter
scale waves and, since we can apply filters only in the flight direction, we cannot isolate the components relevant for the comparison with GLORIA.

9.3.2 Comparison to aircraft in situ data: Scandinavia case

On 28 January 2016 HALO performed a stacked-level flight over southern Scandinavia (see Fig. 9.8). From the base in Kiruna HALO first flew southward. At around 61° latitude HALO turned westward (segment 1). At the end of segment 1 HALO made a turnaround and flew back (segment 2). Turning south, HALO climbed to about 12.5 km altitude and flew a parallel leg (segment 3) and after climbing to about 13.2 km altitude flew segment 4. In addition, there is a flight of the Falcon aircraft in parallel to segment 1, but at about 10 km. The timing of this flight was chosen in a way that the two planes flew almost directly on top of each other. Thus, segments 1, 2, 4 and the Falcon flight probe a vertical cross section at 61° latitude, and section 3 is designed to view this cross section with GLORIA. Particularly suited for comparison is the retrieval based on GLORIA data taken on segment 3 viewing the Falcon flight (Fig. 9.9) and the retrieval of GLORIA data taken on segment 4 viewing segment 3 (Fig. 9.10). In both cases, there is a pronounced GW structure in the in situ observations, and GLORIA views the flight leg with the in situ observations from above.

The temperature residuals shown in Fig. 9.9a are in good agreement. All three data sets, MA6 (i.e., the Falcon in situ probe), GLORIA and ECMWF display the superposition of a large-amplitude, longer scale wave with ~300 km wavelength and shorter
Figure 9.8: Flight segments of the Scandinavia flight on 28 January 2016 in terms of a map (panel a) and as an altitude profile (panel b). Time is given in UTC. For further description see text.
Figure 9.9: Comparison of temperature and winds from various data sets, Scandinavia, 28 January 2016. Data are for the position of the Falcon flight viewed from segment 3. Shown are perturbations from MA6 (light green), smoothed MA6 data (dark green), ECMWF (light blue), S3D reconstruction from ECMWF (dark blue), GLORIA (orange) and S3D reconstruction from GLORIA (red). Panel a shows perturbations for temperature, panel b for zonal wind, panel c for meridional wind and panel d for vertical wind. For discussion see text.
scale waves of ~100 km wavelength with smaller amplitudes (about 0.5 to 1 K amplitude). The S3D fits capture the larger scale wave, but somewhat underestimate the amplitude. Zonal winds are in good agreement. The deviations of the GLORIA S3D reconstruction are consistent with the deviations in temperature. Meridional winds are generally small and very little is from the large scale GW. Vertical winds from MA6 are again dominated by short-scale fluctuations, particularly eastward of 10° longitude. After smoothing, the vertical winds from MA6 are consistent.

Figure 9.10: As Fig. 9.9, but for segment 3 (in situ) viewed from segment 4.

For segment 3 (Fig. 9.10) temperature structures agree very well with exception of an additional maximum in the BAHAMAS data at 13° longitude. This additional maximum is visible as an additional zonal wind maximum at around 11° longitude. Meridional winds are small and generally consistent but hard to distinguish from other signatures. Vertical wind fluctuations show large short-scale fluctuations eastward of 8° longitude. The smoothed BAHAMAS data agree well with the large scale wave patterns from ECMWF and S3D reconstructions from GLORIA and ECMWF.
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The two legs of the Scandinavia flight are hence a quantitative corroboration of the polarization for a large amplitude GW.

9.4 Comparison to dropsonde data

Dropsondes can as a rule not be released over inhabited areas. Therefore from the two GWEX flights dropsondes were released only for the Iceland crossing. Six dropsondes were released at the positions marked in Fig. 9.11. The short traces inside the dots show that these soundings are almost vertical, i.e. except for very short horizontal wavelengths the vertical wavelength can be well estimated.

Dropsonde operations data are available on the HALO data base. For this study, analyses were performed and provided by Sonja Gisinger and Stefan Kaufmann (DLR-IPA). The evaluation follows a well established method.

For the wave analysis the dropsonde data were interpolated onto an equidistant vertical grid with 50 m resolution. The wave perturbations $u'$, $v'$, and $T'$ were calculated as differences between the actual quantities and the respective background profiles. The background profiles were determined by a second-order polynomial fit of $u$, $v$ and $T$ in the troposphere (tropopause region introduces artificial outliers). Additionally, a 5 km running mean was removed from the perturbation profiles and added to the background (Lane et al., 2000, 2003). Hodograph analysis and Stokes parameters are used to describe essential parameters of the gravity waves retrieved from the perturbation wind components ($u'$, $v'$) and from $T'$. The associated techniques are well documented and described, e.g. by Vincent (1984), Eckermann and Vincent (1989), Eckermann (1996), Vincent et al. (1997), and Murphy et al. (2014).

Since this method does not work across the tropopause and since the stratospheric portion of the dropsonde measurements is very small, here we provide analyses for the troposphere only.

For the further discussion we focus on dropsonde 4. Vertical profiles of temperature, zonal wind and meridional wind are shown in Fig. 9.12. The tropopause is around 9 km altitude. The solid lines indicate the measurements, dashed lines give the smooth background. The resulting residual profiles are shown in Fig. 9.13. The dominant wave mode is isolated by a sinusoidal fit to the vertical perturbation profiles (red lines). In general the fit captures the variations, but there are phase shifts for temperature and meridional winds above 6 km, indicating that some altitude dependent analysis method is required. The fit parameters result in a smooth hodograph and the corresponding wave properties (Fig. 9.14). In contrast, the measured data show a superposition of a multitude of GWs with different scales. More sophisticated methods would be needed for an attempt to separate these waves. This, however, is not standard technique and beyond the scope of this study.
Figure 9.11: Map showing the locations of the dropsonde releases over Iceland, 25 January 2016. The map gives smoothed background temperatures from ECMWF. The black line gives the HALO flight path. The circles mark the dropsonde releases, color gives the dropsonde temperature at 11.5 km. The black lines inside the dots show the horizontal projection of the dropsonde fall.

Figure 9.12: Vertical profiles of temperature, zonal wind and meridional wind for dropsonde 4. The solid lines indicate the measurements, dashed lines give the smooth background.
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Figure 9.13: Vertical profiles of residuals from a smoothed background for temperature, zonal wind and meridional wind for dropsonde 4. The black lines indicate the observations, red lines give the wave fit through the profiles.

Figure 9.14: Hodograph of the wind profiles of dropsonde 4. The left panel shows the observations, the right panel the fit results.
Figure 9.15: Comparison of dropsonde 4, ECMWF and GLORIA profiles.

We have also made the attempt to use the smoothed ECMWF data for a common background removal, same as for the BAHAMAS data in Section 9.3.2. The results are shown in Fig. 9.15.

Considering all dropsondes, DS1 to DS3 show, like DS4, a superposition of many different scales. DS5 and DS6 correspond to very short horizontal wavelengths (28.5km and 48.2km, respectively) invisible to GLORIA. Their propagation direction are at very different angles.

The complex wave situation in the Iceland case makes comparison of GLORIA and dropsonde data difficult. In order to have a meaningful direct comparison of residual profiles, similar to the BAHAMAS data, one would need a larger overlap of the observation altitudes. More complex techniques for comparison would need to be employed, such as for instance analyzing the dropsondes for a superposition of waves, projecting the wave properties across the tropopause and comparing the waves through the GLORIA observational filter. Another approach might be comparison via e.g. high resolution modelling. Both approaches would require substantial development effort and are beyond the scope of this study. It should be noted that dropsondes and GLORIA are not per se difficult to compare. For instance, for PGS flight 18 on 6 March 2018 some clear monochromatic waves can be identified in the hodographs.
9.5 Vertical displacements visible in trace species

9.5.1 Vertical displacements visible in trace species: Iceland

As discussed in Chapter 2, Section 2.2 vertical air parcel displacement associated with gravity waves results in adiabatic heating and cooling. Therefore there is a direct, quantitative relation between the highest displacements of the air and the cold fronts and the largest downward displacements and the warm fronts given by Equation (2.6). This is indicated by the displacement of the conserved quantity, i.e. in the displacement of surfaces of constant potential temperature. Furthermore, in the absence of wave breaking, up- and downward transport of air masses along the phase lines represents a reversible transport process for trace species, which usually are transported in the horizontal on surfaces of constant potential temperature. Thus, vertical air mass displacement within gravity waves should be directly observable as oscillations in trace gas field, if a tracer shows a sufficiently large vertical gradient and a horizontally homogeneous distribution on large scales. Isolines of the trace gas mixing ratio are expected to follow the wave-induced deformations of isentropes in the vertical domain.

However, in reality, tracer fields are modulated by synoptic scale atmospheric motion, horizontal transport phenomena, mixing processes and the previous history of air masses including wave perturbations and breaking on all scales. This poses the question, particularly in the UTLS, whether wave signatures can be identified reliably in trace gas fields and used to derive polarization relations. In the context of infrared limb observations, this would allow to further constrain properties of gravity waves identified in the temperature field.

To address this question, simulations by the high-resolution forecast model ICON in combination with the ART module (Schröter, 2018; Schröter et al., 2018) for the Arctic winter 2015/16 by Diekmann (2017) are used to investigate gravity wave signatures in the trace gas fields associated with the GLORIA flights on 25 and 28 January 2016. As water vapor can be derived from infrared limb observations and shows a strong tropospheric gradient, we investigate ICON-ART specific humidity as a tropospheric tracer. On the other hand, ozone is also accessible to infrared limb observations, and therefore we analyze ozone simulated by ICON-ART as a stratospheric counterpart to test whether wave signatures and polarization can be inferred.

Figure 9.16 shows the flight path of the Iceland flight together with residual temperature, specific humidity and ozone simulated by ICON-ART at 12 km and 10 km altitude. Here, residual temperature is calculated by subtracting the moving mean temperature within a window of 3° latitude and and 7° longitude (corresponding with about 330 km times 330 km for the Iceland case) from the local temperature at each grid point. In the area covered by the hexagon flight pattern, the temperature fingerprint of the mountain wave discussed before is clearly visible. Cold and warm phase fronts are aligned in a south-west-west to north-east-east direction at both altitude levels in Figures 9.16a
Figure 9.16: ICON-ART horizontal distributions of residual temperature (a,d), specific humidity (b,e) and ozone (c,f) at 12 km and 10 km altitude, respectively, for the Iceland flight on 25 January 2016. Green dashed lines indicate the positions of the vertical cross sections shown in the following Fig. 9.17.
Figure 9.17: ICON-ART vertical distributions of temperature (a,d), specific humidity (b,e) and ozone (c,f) at 66°N and 15°W, respectively, for the Iceland flight on 25 January 2016. Black dashed lines indicate potential temperature. Green dashed lines indicate intersection between the meridional (upper row) and zonal (lower row) cross sections.

and d. Consequently, strong temperature modulations are found in meridional direction.

The horizontal distribution of specific humidity shown in Fig. 9.16b at 12 km shows no characteristic signature: since stratospheric air masses are found here, water vapor shows hardly any gradient. At 10 km (Fig. 9.16e), enhanced water vapor in the southeastern corner of the hexagon coincides partly with a cold front identified in Fig. 9.16d and may be the consequence of uplifted air masses. However, no clear assignment is possible, as the structure overlaps with an extended ridge in the tropopause characterized by enhanced specific humidity eastward of 10°W. However, in the horizontal ozone distributions shown in Figures 9.16c and f, enhanced/reduced mixing ratios coincide with the warm/cold phase fronts found in panels a and d and are likely an indication for adiabatic downward/upward transport.

From the alignment of the wave signatures Fig. 9.16 it is obvious that stronger modulations of the temperature and trace gas fields on a defined horizontal length have to be expected in the vertical domain in meridional direction than in zonal direction. Fig. 9.17 shows zonal (panels a,b,c) and meridional (panels d,e,f) vertical cross sections of temperature, specific humidity and ozone together with isentropes. As expected, in zonal direction only weak deformations of isentropes are found. In contrast,
in meridional direction, wave-signatures are found in the isentropes around 66°N as the consequence of the temperature modulations in the temperature field. In Fig. 9.17e and f, isolines of specific humidity and ozone follow the isentropes well between 64°N and 67°N below 13 km. Here, the vertical displacement induced by the mountain wave can be clearly deduced from the modulations in the tracer fields. However, the specific humidity isolines show a significantly different pattern than the 320 K isentrope around 68°N due to a local folding event, and Fig. 9.17b shows around the same isentrope that specific humidity does not necessarily follow the isentropes due to deformations and folding of the tropopause region. The simulated ozone distribution (Fig. 9.17c and f) furthermore shows significantly different patterns when compared to the isentropes above about 12 to 13 km.

In Fig. 9.18 the GLORIA observations along segment 6 from 1D retrievals are shown. In contrast to ICON ozone is not increasing monotonously to higher altitudes. There is a maximum around 11 km altitude with a lot of fine structure. This may be due to filamentation by large horizontal transport or small scale GWs. The lower edge follows quite well the 330 K isentrope. In particular around 9:30 UTC we find an upward displacement in both potential temperature and ozone. The difference to the lowest point of the isoline around 9:40 is about 400 m. GLORIA ozone indicates somewhat larger values than ECMWF potential temperature. The value is in reasonable agreement to a displacement amplitude (i.e. half of a peak to peak variation) of 250 m obtained from temperature amplitudes via Equation (2.6).
Figure 9.19: ICON-ART horizontal distributions of residual temperature (a,d), specific humidity (b,e) and ozone (c,f) at 12 km and 10 km altitude, respectively, for the Scandinavia flight on 28 January 2016. Green dashed lines indicate the positions of the vertical cross sections shown in the following Figure.

9.5.2 Vertical displacements visible in trace species: Scandinavia

For the Scandinavia flight shown in Fig. 9.19, the local warm and cold fronts probed by the flight are aligned in north to south direction. Residual temperature is calculated using running means as for the Iceland flight case. Strong temperature modulations are found both at 12 km and 10 km. While specific humidity at 12 km in Fig. 9.19b again shows no signature due to the absence of a stratospheric gradient, cold phase fronts in Fig. 9.19d are reflected by enhanced specific humidity due to upward transported air mass in Fig. 9.19e (vice versa for warm phase fronts). At 12 km, the simulated ozone distribution (Fig. 9.19c) furthermore shows a south-to-north oriented pattern of enhanced ozone east of 10° E. The warm front, however, is centered around 10° E. Thus, ozone and temperature are not in the phase relation expected for adiabatic transport. Gravity waves also induce a displacement perpendicular to the wave vector. However, this structure is far too large in the south-north direction to be caused by the GW. In Section 9.2.3, Fig. 9.3, we have seen a meridional wind pattern not related to the GW. Therefore, in the meridional winds as well as in the trace species there is indication of a further dynamical process other than the strong GW observed here. At 10 km altitude (Fig. 9.19d to f) and 60° N, there are west/east of 10° E cold/warm temperature residuals (panel d), increased/reduced water vapor values (panel e) and
reduced/increased ozone mixing ratios (panel f). These patterns are restricted to the region of the largest wave signatures in the vicinity of the HALO flight pattern. The consistency of two trace species and the horizontal match of the pattern are evidence that at 10 km altitude the patterns are caused by the vertical displacement in the GW.

In the vertical cross sections of the ICON-ART data, the wave signatures can be clearly identified in the isentropes in zonal direction in Figures 9.20 (a,b,c). In the meridional direction (e,d,f) some wave signatures are visible around 64° N, where the cross section intersects the waves above the Norwegian coast. In the zonal cross section, the strong vertical gradient in specific humidity follows the 320 K isentrope well (Fig. 9.20b). However, above, the specific humidity isolines significantly differ from the pattern of the isentropes. Between 5° E and 15° E the isolines follow the isentropes of 320 K and 340 K. At 360 K (12 km) and above the unidentified dynamical process dominates.

Fig. 9.21 compares the ICON simulations to a GLORIA 2D cross section of ozone. Instead of a continuous increase of ozone, GLORIA rather shows a layer of enhanced ozone around 11 km sandwiched between lower ozone mixing ratios above and below. Horizontally this layer is structured, likely by horizontal transport. However, the lower edge of the layer roughly follows the undulations of the 330 K isentrope from
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ECMWF and also the upper edge shows a wave-like structure. It is not very certain to disentangle this wave displacement from the horizontal structure, but in a tentative attempt we find a peak-to-peak variation of 800 m, corresponding to an amplitude of 400 m. Using fitted temperature amplitudes and Equation (2.6), we expect a displacement amplitude of 250 m. Given the uncertainty in the layer-height determination and the fact that the fitted amplitudes underestimate the local temperature variations this is in reasonable agreement.

9.6 Summary

The polarization relations are based on WKB theory of small disturbances. For large amplitude waves in a shear environment these conditions are not fully justified. Theoretical investigations suggest that they should hold in a wider range of conditions. The combination of GLORIA and HALO and Falcon in situ instrumentation provide one of the very rare opportunities to test the polarization relations in single case studies from observations.

For the Iceland flight good consistency between temperatures and meridional wind is found. This is further evidence that all isolated temperature structures are GWs. For the Scandinavian flight we find good agreement both for the inferred zonal wind (parallel to the wave vector) and the vertical wind. This is a direct experimental support for the validity of the polarization relations also for large amplitude waves.

The distributions of trace species are dominated by the general vertical gradient and horizontal transport. Nonetheless, in the presence of strong GWs these gradients or layers of trace species are visibly displaced. In a case study approach which can judge the relative importance of horizontal transport and vertical displacement in a certain region, a check on the consistency of temperature perturbation and tracer dis-
placement can be made. This, however can only confirm the size of the temperature amplitude and the nature of adiabatic transport (versus diabatic processes such as e.g. convection in the troposphere), as any adiabatic vertical transport leads to the same result. In contrast, the polarization for the winds are GW-specific and thus confirm the nature of the process as really being a GW.
10 Conclusion and Outlook

10.1 Summary of the study

Gravity waves (GWs) are small scale and mesoscale atmospheric waves, which play an important role in the dynamics of the middle atmosphere. They are difficult to represent in models and many open science questions remain in this field. In particular, better quantitative information about the global distribution is essential to improve climate projection and medium and long range weather forecast.

To close this gap in our knowledge, global observations are mandatory. One proposed concept is a space-borne infrared limb imager (IRLI) that observes 3-D temperature distributions along the orbital track. This 3-D information can be used to infer the 3-D wave vector and GW amplitude, and based on this GW momentum flux. The concept was studied extensively in preparation of EE7 mission selection for the PREMIER proposal. The technical feasibility of the instrument and the required retrievals were investigated in dedicated studies (e.g. PREMIER — Consolidation of Requirements and Synergistic Retrieval Algorithms; (ESA, 2012)). A further study (Gravity Wave Observations From Space; (Preusse et al., 2012)) showed the feasibility to observe GW momentum flux based on simulated data. However, a demonstration based on real data was missing.

For this purpose, ESA funded the GWEX study. The aim of this study was to perform GW observations with the airborne IRLI GLORIA on board of HALO. In this way for a first time 3-D tomographic reconstructions of mesoscale GW events in the lower stratosphere were taken.

The winter 2016 was an unusual one with a particularly cold polar stratosphere. The vortex was, in general, very strong, but also had a wider diameter: the polar night jet (PNJ) was shifted southward compared to the average of previous 10 years. In particular in the observation altitude of GLORIA around 10 km this resulted in unusually low average wind velocities above northern Scandinavia. Therefore our GW observations were carried out farther south.

Two gravity wave events were observed. On 25 January 2016 a mountain wave above Iceland was reconnoitered. On 28 January 2016 GWs excited by strong winds approaching the southern part of Scandinavia were investigated. For both events tomographic retrievals were performed and the 3-D wave structures were investigated. Consistent pictures of these GWs and their excitation could be inferred. By initializing
ray-tracing simulations with the inferred wave parameters, orography is identified as an important source in both cases. It should be highlighted here that few observation techniques offer the possibility of a full wave characterization required for this type of model study. Compact results can only be reached, if the wave parameters are determined with high accuracy (cf. the sensitivity study on the impact of the ray-tracing launch parameters performed by Krisch et al. (2017) in the frame of GWEX).

For the Iceland case, all observed GWs can be back-traced to orography. The wave vectors turn in a rotational wind shear and the wave is propagating very obliquely several thousand kilometers already below 20 km altitude. Up to about 14 km, this oblique propagation is immediately visible in the 3-D tomographic reconstruction of the observed wave field. In the mid-stratosphere, these waves merge into a large field of GWs from different source regions. The AIRS instrument is capable to observe these waves at high altitudes, in principle, but due to the complexity of the situation, it is infeasible to pinpoint the propagated Icelandic wave.

For the Scandinavia flight, orography is also identified to be the most important source. However, spontaneous imbalance above the North Sea contributes as well. From the observation altitude, waves propagate slightly northward but remain over Scandinavia. Analysis of AIRS data shows enhanced amplitudes mainly over the southern tip of Scandinavia. This is due to the fact that only the background wind over Southern Scandinavia is sufficiently strong such that vertical wavelengths increase to values well inside the visibility filter of AIRS. The visibility filter does not only reduce the amplitude but also modifies the predominant vertical wavelength of the observed wave. Improved matches may be found in stronger background winds: despite the fact that in January 2016, on average, the wind was exceptionally strong, this was not the case for this particular day.

The combination of GLORIA’s 3-D capability with in situ observations of winds offered the rare opportunity to test the polarization relations. We approached this task by first investigating ECMWF model data. These have the advantage to provide global 3-D fields on synoptic sampling resolving the larger part of the mesoscale GW spectrum. An essential step in any GW analysis is the separation into GW perturbations and background. We found that the usual separation into global scale Rossby waves and mesoscale GWs applied, for instance, to satellite data in the stratosphere (Fetterer and Gille, 1994; Ern et al., 2006, 2018) still works, but requires adjustments to the employed filters: the lower the altitude and the closer to the troposphere the higher are the wavenumbers which need to be assigned to the background. Instead of a wavenumber 6 sufficient in the stratosphere we had to use wavenumber 18 and even 36 for the horizontal winds. This implies that gravity wave perturbations in temperature and vertical winds are easier to separate from the background than perturbations in horizontal winds.

For the model data, a good agreement between directly inferred GW residuals and GW perturbations constructed from S3D temperature analyses based on the polariza-
tion relations is obtained. This confirms the application of the polarization relation for calculating momentum fluxes and the suitability of the analysis tools. In the Scandinavia case, the background removal may have only partially worked. Therefore, also other processes aside the effect of GW are still visible in the wind residuals.

Progressing to real data, we face the following difficulties: a common background removal has to be applied in order to keep the data comparable. Applying different detrending techniques would imply different observational filters and thus make the comparison problematic from the beginning. Inherently to the method, GLORIA retrievals and ECMWF models resolve only the mesoscale part of the spectrum. In situ measurements contain a much wider part of the spectrum including short scales down to $O(1\ \text{km})$. We have filtered the data accordingly, but due to the 1-D nature of the in situ data, this filtering takes only into account the apparent wavelength projected to the observation track.

In the Iceland case, we can achieve consistency between temperatures and meridional wind (the main horizontal wind component of the mesoscale GWs) measured by the HALO in situ instrumentation BAHAMAS. This confirms that the observed mesoscale structures obey the GW polarization relations and are indeed GWs. Vertical winds are dominated by short scale waves and since obviously many different propagation directions contribute, along-flight-path filtering cannot successfully suppress them. The drop sondes were released in the Iceland flight from a relatively low flight altitude, which was optimized for the match between GLORIA and BAHAMAS. This makes the overlapping altitude range in the stratosphere small. In general, the presence of many small scale waves with different propagation directions prevents from the direct comparison of the GW perturbation of these two techniques in this case.

For the Scandinavia case, we compared to BAHAMAS and the in situ sensors MA6 of the Falcon aircraft. In this case both longer and shorter scale waves were aligned chiefly opposite to the background winds and had a propagation direction due westward. Though we also found shorter scale fluctuations dominating the vertical wind, these could be filtered. The comparison of the winds deduced from temperature S3D fits via the polarization relation confirms the use of the polarization relation for both zonal and vertical wind.

We also investigated the displacement of trace species (water vapor and ozone) by GWs. As expected, GWs do displace the tracers. In case studies, we can find consistency between the temperature and trace species structure. For the interpretation of the trace species distribution, the GW signal is important. However, dominant is still the horizontal transport. Therefore, one cannot infer GW properties from tracer distributions alone.
10.2 Lessons learned for campaigns

What did we learn for future campaigns? Despite the fact that we reached all major aims and gained new scientific insights there is always room for improvement.

In order to follow the oblique propagation of the Iceland wave upward, we would have needed an upward looking observation of temperature and/or winds on HALO. DLR-IPA will install the ALIMA lidar system which is capable of observing temperature and along-line-of-sight winds with good vertical and temporal resolution in the entire stratosphere and up to the mesopause.

For better matches with AIRS, we need more-isolated sources and stronger background winds. Both is the case for South America and the Antarctic Peninsula. A campaign with a base in South America is in preparation.

In setting up campaign planning tools, we anticipated oblique propagation, but not to the degree as observed in the Iceland case. Also the temporal evolution of the winds needs higher attention.

For comparison with GLORIA, dropsondes should be released from as-high-as-possible flight altitudes.

10.3 Implications for a space-borne infrared limb imager

The chief aim of GWEX was to demonstrate the suitability of IR limb imaging for GW research. This aim was clearly reached. Two GW events were observed. The physical interpretation and the match of polarization relation proof not only the maturity of the technique, but also the very high degree of accuracy reached.

What changes when bringing an IRLI into space? What are the disadvantages and advantages in comparison to the airborne GLORIA?

To start with the disadvantages: the spatial resolution will be worse. With GLORIA we reach a spatial resolution of 200 m in the vertical and around 20 km in the horizontal for full angle tomography (FAT). The PREMIER studies concluded that a point spread function of 800 m is the best resolution which can be realistically reached from a stable low Earth orbit (600 km to 800 km flight altitude). This way we will not be able to resolve some waves in the low stratosphere, but wavelengths become larger at higher altitudes. Upward of 20 km altitude this restriction is not very important anymore. Furthermore ≈1 km is an order of magnitude better than the 10 km vertical resolution of nadir sounding, the only other 3-D-capable space-borne observation technique. The horizontal resolution will be similar to that of the Scandinavia case. In flight direction 2-D tomography will be applied and a spatial resolution of 50 km can be reached (cf. previous PREMIER studies CORSA and GW Observations From Space
Across-track, the resolution depends on the pixel size and will be of the order of 25 km. This is sufficient also to resolve the smaller scale waves (wavelength of $\approx$100 km) observed in the Scandinavia case.

The advantages of a space-borne IRLI compared the airborne demonstration are: Most important, we will gain global coverage. Second, in the stratosphere GWs can be separated more readily from the background than in the UTLS. The amplitude of the GWs grow exponentially with increasing altitude. Furthermore, at these altitudes global scale dynamics is dominated by Rossby waves (extratropics) and e.g. Kelvin waves (tropics) which on the sphere are periodic and can be fully captured by a low Earth orbit satellite. Finally, covering a larger altitude interval and having global coverage one will find cases where one can observe the upward propagation of waves self-consistently from different orbits of the IRLI, instead of mixing different instruments and different observational filters as in the current study.

The IRLI is up to now the only technique where simulation studies and, now, airborne demonstration prove the ability to gain global distributions of GW momentum flux and its direction with the required accuracy.
A HALO database

The HALO data base is a web platform of a data retrieval and long-term archiving system. It was established to hold and manage a wide range of data based on, or related to observations of the HALO research aircraft. The HALO-DB may also be used for sharing data of scientific missions involving other aircraft and in-situ instruments. (HALO database)

Figure A.1: Overview of PGS flights.

The POLSTRACC data is subject to a data agreement restricting access during a fixed time frame following the campaign, which is over at the current point in time.
We uploaded cross-section data for all flights for both the dynamics and chemistry mode measurements of GLORIA. For tomographic retrievals, only data retrieved for publications has been uploaded due to the high effort of providing and validating the data. Figure A.1 shows an overview over the flights of the POLSTRACC campaign together with the dates of the flights.

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This table shows processed data and availability. The last column contains hyperlinks to the data sets on HALO database. The notation of “FZJ 1-D” provides links to the data sets of the FZJ retrievals for dynamics mode data. The notation of “KIT 1-D” provides links to the data sets of KIT cross-section retrievals. The notation of “FZJ 3-D” refers to 3-D tomographic data sets derived from tomographic dynamics mode measurements. The two flights highlighted in bold letters mark the two flights of the GWEX sub-campaign.

The first four flights are missing, partially due to the technical nature of the flights and partially due to instrumental problems. “FZJ 1-D” processing is missing for flight 05 due to small number of available profiles and for flight 08 due to high contamination of spectra by polar stratospheric clouds. “KIT 1-D” processing is missing for flights 15 and 18 due to small number of measured chemistry mode spectra. KIT processing has been described in detail by Johansson et al. (2018). Tomographic temperature volumes are available for flights 10 and 11. Flight 10 offers two tomographic retrievals of the same volume, one derived by full-angle-tomography from the hexagonal pattern surrounding the volume and one from the limited-angle-tomography from the crossing of the volume.
B Bibliography


HALO database. HALO database, 2017. URL [https://halo-db.pa.op.dlr.de/](https://halo-db.pa.op.dlr.de/) last access: 1 December 2017, uploaded on: 1 December 2014.


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