

Response to the comments of reviewer #2,

we are thankful to the reviewer, whose comments helped us to improve the paper. We have revised the paper according to the remarks, and hope that we sufficiently responded to each concern.

**SC 2:** We already discussed most of the comments of reviewer #2 in detail and now shortly summarize most of the points.

In the following we are using the abbreviations **P** and **L** for page and line referring to the revised paper.

We attached the revised version of the paper. The blue color refers to the comments of reviewer #2.

**The authors should include a table summarizing the different experiments.**

We insert a table including the position as well as the positive and negative maxima of the zonal and meridional GW drag and heating by GWs.

**P6**

**The reviewer found it a bit confusing that the Gaussian GW hotspot is shown in Figure 2 and the comparison between the ‘used’ simulations and the Gaussian is later presented in Figure 8. It is suggested to present this type of sensitivity in the experimental description section and later just refer to the ‘main’ model experiments.**

We added more explanations to the description of Figure 2 and the new Table 1, and clearly stated that we will come back to the Gaussian-shaped hotspot later.

**P7L2-3, P7L17-21**

**Please clarify whether the altitudes shown in the Figures correspond to the pressure grid or whether the altitudes are computed from the geopotential and converted into a geometric altitude. This simplifies some comparisons to observations results.**

We added an explanation of the logarithmic pressure height in the revised paper.

**P3L18-22**

**Another comment of the reviewer concerns the implementation of the GW drag for the different latitudes. As the drag scales also with the atmospheric air mass that is affected by the drag, it might be mentioned that at higher latitudes essentially less drag is exerted to the atmosphere as the GW drag volume scales with latitude. Or with other words the atmospheric mass that is affected by the drag decreases with latitude. This might need some more discussion or should at least be mentioned in the interpretation of the results.**

We added a note on this in section 2 and listed some arguments explaining why we did not weight the artificial GW forcing with the latitude.

**P7L22-31**

**Some Figures (3,4,6 and 7) need an improvement of the quality.**

We improved the resolution of these figures.

**P8, P9, P11, P12**

**Page 3: line 18: . . .vertical resolution 2842 km should be 2842 m**

We are using the ‘.’ for the separation of decimal numbers.

# Effect of latitudinally displaced gravity wave forcing in the lower stratosphere on the polar vortex stability

Nadja Samtleben<sup>1</sup>, Christoph Jacobi<sup>1</sup>, Petr Pišoft<sup>3</sup>, Petr Šácha<sup>2,3,4</sup>, and Aleš Kuchař<sup>3</sup>

<sup>1</sup>Institute for Meteorology, Universität Leipzig, Stephanstr. 3, 04103 Leipzig, Germany

<sup>2</sup>Institute for Meteorology, Universität für Bodenkultur Wien, Gregor-Mendel-Straße 33, 1180 Vienna, Austria

<sup>3</sup>Department of Atmospheric Physics, Faculty of Mathematics and Physics, Charles University, V Holesovickach 2, 180 00 Prague 8, Czech Republic

<sup>4</sup>EPhysLab, Faculty of Sciences, Universidade de Vigo, Campus As Lagoas, s/n, 32004 Ourense, Spain

**Correspondence:** Nadja Samtleben (Nadja.Samtleben@uni-leipzig.de)

**Abstract.** In order to investigate the impact of a locally confined gravity wave (GW) hotspot, a sensitivity study based on simulations of the middle atmosphere circulation during northern winter was performed with a nonlinear, mechanistic, global circulation model. To this end, for the hotspot region we selected a fixed longitude range in the East Asian region (120°E-170°E) and a latitude range from 22.5°N-52.5°N between 18 km and 30 km, which was then shifted northward in steps of 5°. For the southernmost hotspots, we observe a decreased stationary planetary wave (SPW) 1 activity in the upper stratosphere/lower mesosphere, i.e. less SPWs 1 are propagating upwards. These GW hotspots are leading to a negative refractive index inhibiting SPW propagation at midlatitudes. The decreased SPW 1 activity is connected with an increased zonal mean zonal wind at lower latitudes. This in turn decreases the meridional potential vorticity gradient ( $q_y$ ) from midlatitudes towards the polar region. A reversed  $q_y$  indicates local baroclinic instability, which generates SPWs 1 in the polar region, where we observe a strong positive Eliassen-Palm (EP) divergence. As a result, the EP flux is increasing towards the polar stratosphere (corresponding to enhanced SPW 1 amplitudes), where the SPWs 1 are breaking and the zonal mean zonal wind is decreasing. Thus, the local GW forcing is leading to a displacement of the polar vortex towards lower latitudes. The effect of the local baroclinic instability indicated by the reversed  $q_y$  also produces SPWs 1 in the lower mesosphere. The effect on the dynamics in the middle atmosphere by GW hotspots which are located northward of 50°N is negligible because the refractive index of the atmosphere is strongly negative in the polar region. Thus, any changes in the SPW activity due to the local GW forcing are quite ineffective.

## 1 Introduction

During winter the dynamics of the middle atmosphere are mainly dominated by the polar vortex, which develops due to the lack of incoming solar radiation, and which is modified by the impact of atmospheric waves with different spatial and temporal scales (Douville, 2009). The most important characteristic of atmospheric waves is their ability to transport and deposit energy and momentum. In particular, gravity waves (GWs) mainly developing in the troposphere distribute energy and momentum throughout the whole atmosphere thereby maintaining the circulation and the thermal structure of the upper atmosphere (Fritts

and Alexander, 2003). They also contribute to turbulence and mixing between all vertical layers. Their most important sources are orography (Smith, 1985; Nastrom and Fritts, 1992), convection (Tsuda et al., 1994), **jet sources** (Plougonven and Zhang, 2014) or spontaneous adjustment processes (Fritts and Alexander, 2003). Strongly depending on the phase speed  $c$  and the background wind  $u$ , GWs are able to propagate into the middle atmosphere. Due to the exponentially decreasing density of the atmosphere the GW amplitude is exponentially increasing with height **if the GWs are propagating conservatively under background conditions that are constant with height**. Usually, the GW spectrum is already saturated in the stratosphere, which means that GW amplitudes cannot grow anymore and, according to the linear theory, partly break. This effect is the stronger the closer their phase speed  $c$  is to the background wind  $u$ . If  $c$  is equal to  $u$ , the GW encounters its critical line and cannot propagate anymore (Lindzen, 1981). Thus, mainly GWs propagating into the opposite direction than the background wind are usually observed in the middle atmosphere. Also GWs being faster than the background wind are able to propagate but they are mostly filtered out by the strong polar-night jet **at the latest, when  $c$  becomes equal to  $u$ . In the mesosphere GWs, which are propagating in the opposite direction to  $u$  saturate and deposit their momentum. For this reason the wind reverses in the upper mesosphere/lower thermosphere (MLT)** (Lindzen, 1981; Holton, 1982). The transfer of energy and momentum by breaking GWs is also called GW drag.

Owing to the variety of sources GWs have a large spatial and temporal variability. To capture the global distribution of GWs, the potential energy ( $E_{pot}$ ), momentum flux (MF), or stability indicators (Pišoft et al., 2018) can be estimated by using satellite data (Ern et al., 2004; Fröhlich et al., 2007; Hoffmann et al., 2013; Schmidt et al., 2016). These numerous observational studies highlight a number of different local GW hotspots, which are mainly generated by orography and convection. The most common GW hotspots are the orographically induced GW hotspots near the Alps (Hierro et al., 2018), the Andes (Llamedo et al., 2009; Alexander et al., 2010; Lilienthal et al., 2017), the Antarctic Peninsula (Moffat-Griffin et al., 2010), the Himalayas (Kumar et al., 2012), the Mongolian Plateau (White et al., 2018), the Rocky Mountains (Lilly et al., 1982) and in the Scandinavian region (Kirkwood et al., 2010). **Typically, satellite observations show a characteristic structure of enhanced GW activity in the subtropical stratosphere that is caused by deep convection over Southeast Asia, America, Africa, or the Maritime Continent in the respective summer season (Jiang et al., 2004; Wright and Gille, 2011; Ern and Preusse, 2012). However, reliable estimates of GW drag from observations are generally difficult. Several methods have been established to derive the GW drag from satellite (e.g. Ern et al., 2014, 2016), or from radar measurements (e.g. Reid and Vincent, 1987). However, uncertainties of these estimates are quite large.**

From model studies there are indications that GWs can already break in the lower stratosphere (LS) (e.g. Plougonven et al., 2008; Constantino et al., 2015), which leads to an additional transfer of momentum and energy in this region. In connection with high PW activity this is highly affecting the stability of the polar vortex and can cause a sudden stratospheric warming (SSW) (Albers and Birner, 2014). This effect has been also observed in satellite measurements showing enhanced GW drag before SSWs (Ern et al., 2016). Thus, an additional GW forcing may lead to a preconditioning of the polar vortex.

**This study is focusing on the role of the zonal position of localized GW breaking areas and their effects on the middle atmosphere dynamics.** It is motivated by the findings of Šácha et al. (2015) who were focusing on the East Asian/North Pacific (EA/NP) region near Japan where they observed a GW hotspot being active during equinoxes and winter solstices. The GWs

are orographically and convectively generated due to the topography directly at the coastline and the warm Kuroshio current. Šácha et al. (2015) analysed the local instabilities by calculating the Richardson number and by analysing reanalysis data and found that the GWs are breaking in this area. Based on these results they simulated the observed Asian GW breaking hotspot with a global circulation model (GCM) and analysed its effect in the middle atmosphere circulation (Šácha et al., 2016). According to previous publications e.g. by Smith (2003), Lieberman et al. (2013), or Matthias and Ern (2018), Šácha et al. (2016) observed a forcing of additional stationary planetary waves (SPWs) due to a longitudinally variable GW drag. We are pursuing this idea by shifting meridionally the EA/NP hotspot keeping its longitude range fixed to get information about its impact on the middle atmosphere at different latitudinal positions. Therefore, the EA/NP GW hotspot is our starting point, from which we are displacing the GW hotspot towards lower and higher latitudes in  $5^\circ$  steps. In section 2 of this paper, we provide a brief description of the GCM and detail the implementation of the GW hotspot within the GCM. In section 3 we describe and discuss the observed effects of the GW hotspots on the circulation of the middle atmosphere by analysing the stationary planetary wave activity and the propagation conditions. Finally, conclusions and outlook are presented in section 4.

## 2 Numerical model experiments

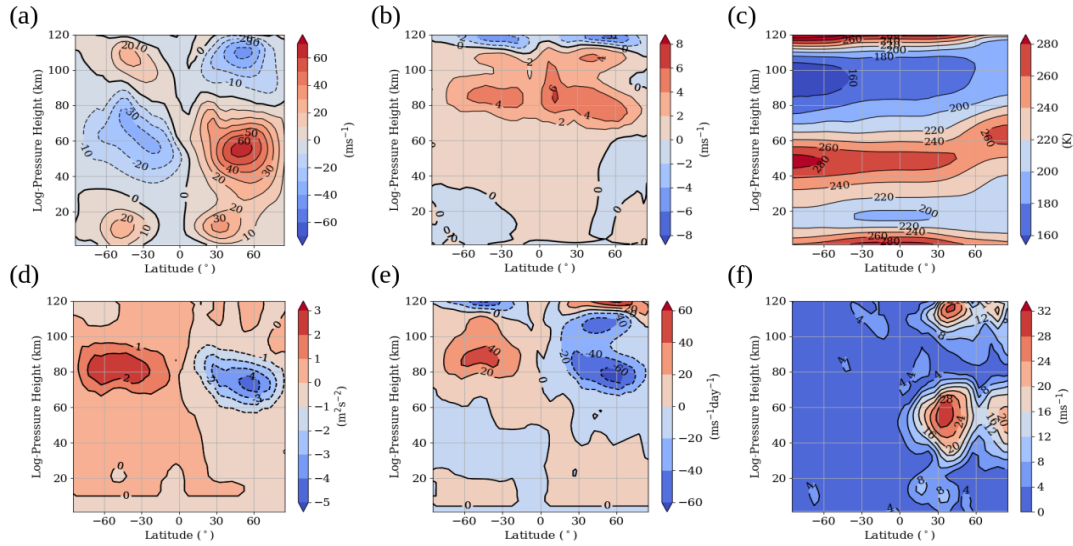
### 2.1 Model description and set up

To investigate the effect of localised GW breaking hotspots in the LS simulations have been performed using the Middle and Upper Atmosphere Model (MUAM, Pogoreltsev et al. (2007)). MUAM is a non-linear mechanistic 3D grid point model, which is an updated version of the global circulation model COMMA-LIM (Fröhlich et al., 2003a, 2007; Jacobi et al., 2006). The model extends in 56 layers up to an altitude of about 160 km in logarithmic pressure height  $z = -H \ln(p/p_0)$  with a constant scale height  $H = 7$  km and the reference pressure  $p_0 = 1000$  hPa. Depending on the temperature profile the used logarithmic pressure height can differ from the geometric height. However, at altitudes below 80 km, this difference is negligibly small. In 110 km the deviation is increasing up to 5 km, whereas the highest logarithmic pressure level of about 160 km may correspond to a geometrical height between 300 km and 400 km. In the lowermost 10 km, zonal mean temperatures are nudged to 2000-2010 mean monthly mean ERA Interim (Dee et al., 2011) zonal mean temperatures to correct the climatology of the troposphere, which is not included in detail in the model (Jacobi et al., 2015; Lilienthal et al., 2018). Furthermore, at 1000 hPa, which defines the lower boundary of the model, SPWs of wavenumbers 1, 2 and 3 are forced, which are extracted from 2000-2010 mean ERA Interim monthly temperature and geopotential reanalysis data. The horizontal resolution of the model is  $5^\circ$  in latitude and  $5.625^\circ$  in longitude and the vertical resolution is 2.842 km. The model solves the primitive equations in flux form (e.g. Jakobs et al., 1986). MUAM includes parameterizations to simulate subgrid processes such as GWs, absorption of solar radiation, or infrared cooling. The absorption of radiation is realized according to Strobel (1986). This parameterization is focused on the absorption processes due to trace gases such as  $\text{H}_2\text{O}$  (absorber in the troposphere),  $\text{CO}_2$  and  $\text{O}_3$  (absorber in the stratosphere). Water vapor and ozone fields are prescribed. The heating rates are calculated by absorption bands representing the wavelength interval in which these trace gases are absorbing the atmospheric radiation. Infrared emission of  $\text{CO}_2$  is parameterized after Fomichev et al. (1998), and ozone infrared cooling in the  $9.6 \mu\text{m}$  band is calculated after Fomichev and

Shved (1985).

GWs are parameterized after an updated linear scheme (Lindzen, 1981; Jakobs et al., 1986) with multiple breaking levels (Fröhlich et al., 2003b; Jacobi et al., 2006). GW amplitudes are included at an altitude of 10 km as zonal mean with a global average of  $1 \text{ cm s}^{-1}$  for the vertical velocity perturbation. This value is weighted by a prescribed zonal mean GW amplitude distribution based on  $E_{pot}$  data obtained from GPS radio occultation measurements (Šácha et al., 2015; Lilienthal et al., 2017). **Although the  $E_{pot}$  data still contain Kelvin waves and other possible wave structures with short vertical wavelengths, which may introduce biases, the GW amplitude distribution is more realistic compared to the hyperbolic tangent function of the latitude, which was used in earlier experiments (Jacobi et al., 2006) and leads to an improvement of the zonal mean GW climatology. It shows maximum GW amplitudes (not shown here) at the equator (convectively generated GWs) and at midlatitudes (orographically induced GWs).** At each grid point 48 waves are induced propagating in eight different directions with six different phase speeds ranging from 5 to  $30 \text{ ms}^{-1}$ .

In this configuration based on January decadal mean (2000-2010) ERA Interim reanalysis data we create a reference simulation with a spin-up period of 270 days, in which the mean circulation is built up and different waves like planetary waves (PWs) and tides are generated. The declination and the ozone and carbon dioxide concentration are fixed to avoid further non-zonal structures being induced besides to the enhanced GW forcing. The declination corresponds to January 15 (referring to the mid of the month) and the ozone and carbon dioxide data are taken from the year 2005 (refer to the mid of the decade). For the analysis, a time interval of 120 days with a temporal resolution of 2 hours after the spin-up period was modeled. Šácha et al. (2016) have already analysed the effect of the Asian hotspot with MUAM by performing a sensitivity study with regard to the strength of the GW forcing in the stratosphere.



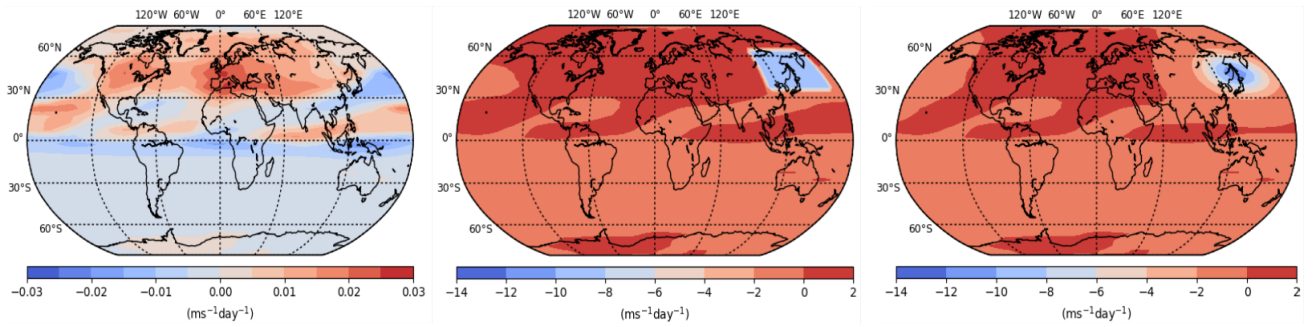
**Figure 1.** January zonal and monthly mean of the (a) zonal wind ( $\text{ms}^{-1}$ ), (b) meridional wind ( $\text{ms}^{-1}$ ), (c) temperature (K), (d) zonal GW fluxes ( $\text{m}^2\text{s}^{-2}$ ), (e) zonal wind acceleration through breaking GWs ( $\text{ms}^{-1}\text{day}^{-1}$ ) and (f) SPW 1 amplitude ( $\text{ms}^{-1}$ ) extracted from the zonal wind of the reference simulation.

Their analysis time period was much shorter and the declination of the sun was different. They also nudged the model zonal mean temperature up to 30 km. In this regard, our experimental set-up might be considered superior to their simulations, especially, because the nudging does not interfere with the implemented GW forcing in this new configuration. We refer to this reference simulation as the ‘Ref’ simulation.

- 5 The state of the middle atmosphere of the Ref simulation can be seen in Fig. 1, which shows the January zonal mean zonal (a) and meridional wind (b), the temperature (c), zonal GW flux (d), the zonal wind acceleration due to breaking GWs (e), and the SPW 1 amplitude extracted from the zonal wind (f) as latitude-height plots. Each parameter is presented up to an altitude of 120 km for the winter and summer hemisphere. The zonal wind in Fig. 1(a) generally reproduces reference climatologies like CIRA-86 (Fleming et al., 1988) or URAP (Swinbank and Ortland, 2003), but the winter mesospheric jet is overestimated by about 10-20  $\text{ms}^{-1}$ . The meridional circulation (Fig. 1(b)) extending from the summer to the winter mesopause has a maximum of 6  $\text{ms}^{-1}$  at about 80 km, which is well reproducing predictions by climatologies (Portnyagin et al., 2004; Jacobi et al., 2009). Temperature (Fig. 1(c)) generally reproduces climatology values. The GW fluxes (Fig. 1(d)) maximize at about 80 km, with a maximum of slightly above  $-4 \text{ m}^2 \text{ s}^{-2}$  on the northern hemisphere (NH) and  $2 \text{ m}^2 \text{ s}^{-2}$  on the southern hemisphere (SH). The corresponding zonal GW drag maximizes at the same altitude with about  $-60 \text{ ms}^{-1} \text{ day}^{-1}$  ( $40 \text{ ms}^{-1} \text{ day}^{-1}$ ) in the NH (SH) and is westward (eastward) directed. The SPW 1 amplitude (Fig. 1(f)) extracted from the zonal wind shows maximum values at the border of the mesospheric jet maximum northward of  $30^\circ \text{N}$  between 50 and 60 km and in the polar region. This fits quite well to observations, but the amplitudes are slightly underestimated due to the overestimated mesospheric jet, filtering some of the SPWs (Xiao et al., 2009).

## 2.2 Experiment description

- 20 In a first experiment, we reproduced the experiment of Šácha et al. (2016) to check if we still get similar results with the slightly modified setup. To represent the Asian GW breaking hotspot in the model we enhanced the GW drag after model day 270, i.e. after the spin up, and run the model for 120 days as in the Ref simulation.



**Figure 2.** Zonal GW drag ( $\text{ms}^{-1} \text{ day}^{-1}$ ) at 26.9 km for the reference (left) and the H3 hotspot simulation as box (middle) and as Gaussian distribution (right) for the last 30 days of analysis. Note the different scaling on the left panel.

Therefore, the zonal ( $GWD_u$ ) and meridional ( $GWD_v$ ) GW drag and the heating due to breaking GWs ( $GWD_T$ ) are modified in the specific region of the observed GW breaking hotspot. In principle, the response to the GW drag would in turn alter the GW propagation and breaking conditions and thus, the GW drag and its distribution. To avoid those feedback mechanisms, the GW parameterization scheme is turned off during the experiments and the model is fed with the GW drag field from the Ref simulation. Only, in the GW hotspot region the GW drag is modified as shown in Tab. 1. We intend to only analyse the steady state impact of the local GW forcing being not influenced by nonlinear effects. Like in Šácha et al. (2016) we located the GW breaking hotspot between 37.5°N-62.5°N and 118.1°E-174.3°E in an altitude range between 18 and 30 km. Note that the geographic positions refer to the model grid points so that at a latitudinal 5° grid the meridional size of the modeled hotspot is 30°. To avoid a total breakdown of the polar vortex and a fundamental change in middle atmosphere dynamics, which was already forced in the study by Šácha et al. (2016), we chose the more moderate case of  $-10 \text{ ms}^{-1}\text{day}^{-1}$  for  $GWD_u$ ,  $-0.1 \text{ ms}^{-1}\text{day}^{-1}$  for  $GWD_v$  and a warming of  $0.05 \text{ Kday}^{-1}$  for  $GWD_T$ . We refer to this simulation as the H3 simulation, as will be described later. The distribution of the  $GWD_u$  of the Ref and the H3 simulation can be seen in the left and middle panel of Fig. 2 at an altitude of about 27 km for the last 30 days of analysis. We are mainly concentrating on the last 30 days of analysis because we are focusing on quasi steady states and are not interested in short term variabilities. The  $GWD_u$  of the Ref simulation is varying between  $-0.025$  and  $+0.02 \text{ ms}^{-1}\text{day}^{-1}$  in the region of the GW hotspots (27.5°N-87.5°N, 118.1°E-174.3°E, 18-30 km). Thus, the maximum value of the H3 simulation ( $GWD_u = -10 \text{ ms}^{-1}\text{day}^{-1}$  in the hotspot) is 500 times larger than the maximum westward (negative) value of the Ref simulation.

Simulation	Abbreviation	Region	Min./Max. $GWD_u$ ( $\text{ms}^{-1}\text{day}^{-1}$ )	Min./Max. $GWD_v$ ( $\text{ms}^{-1}\text{day}^{-1}$ )	Min./Max. $GWD_T$ ( $\text{Kday}^{-1}$ )
Reference	Ref		-0.025/0.02	-0.025/0.01	-0.0006/0.01
Hotspot	H1	27.5-52.5°N	-10	-0.1	0.05
	H2	32.5-57.5°N			
	H3	37.5-62.5°N			
	H4	42.5-67.5°N			
	H5	47.5-72.5°N			
	H6	52.5-77.5°N			
	H7	57.5-82.5°N			
	H8	62.5-87.5°N			
		118.1-174.3°E			
		18-30 km			
Gauss	Gauss		-13.1	-0.13	0.065

**Table 1.** Overview of the mean and maximum values of the zonal and meridional GW drag and heating by GWs for the reference and hotspot simulations as three dimensional box and as Gaussian distribution. The mean and maximum values refer to the region (118.1-174.3°E, 18-30 km and the respective latitude range) of the hotspots.



The H3 mean value (mean  $\text{GWD}_u$ :  $-10 \text{ ms}^{-1}\text{day}^{-1}$ ) is roughly 3300 larger than the one of the Ref simulation (mean  $\text{GWD}_u$ :  $0.003 \text{ ms}^{-1}\text{day}^{-1}$ ) within the region of the EA/NP hotspot. These maximum values of the  $\text{GWD}_u$  as well as those of the  $\text{GWD}_v$  and the  $\text{GWD}_T$  are summarized in Tab. 1 for the Ref and the GW hotspot simulations. In spite of the huge difference compared to the Ref simulation, the zonal GW forcing is moderate in terms of what is estimated from observations ( $40 \text{ ms}^{-1}\text{day}^{-1}$  and more) and from GW parameterizations in this region (Šácha et al., 2018). Concerning the meridional GW drag and the heating due to breaking GWs the maximum (mean) value of the H3 simulation is only 5 (100) times larger than the one of the Ref simulation (not shown here). To investigate possible effects with regard to the position of the GW hotspot we performed a sensitivity study. For this, we kept the longitude ( $118.1^\circ\text{E}$ - $174.3^\circ\text{E}$ ) and altitude (18-30 km) range as well as the zonal extent of  $25^\circ$  fixed, but varied the observed GW hotspot in  $5^\circ$  steps from  $27.5^\circ\text{N}$ - $52.5^\circ\text{N}$  (simulation H1) to  $62.5^\circ\text{N}$ - $87.5^\circ\text{N}$  (simulation H8), while labeling the experiments inbetween by H2 through H7 (see Tab. 1).

To analyze possible effects of the sharp transition zone between the unchanged and enhanced GW drag, additional simulations with a smoothed GW forcing were performed, by using a 3D Gaussian function, with a standard deviation of  $10^\circ$ ,  $22.5^\circ$ , and  $5.684 \text{ km}$  in the zonal, meridional, and vertical direction, respectively. To get the same integral forcing as in the H1-H8 simulations, the size or the intensity of the local GW forcing as Gaussian distribution needs to be adjusted. For our experiments, we mainly increased the strength of the local GW forcing and only slightly increased the size. The maximum values for the  $\text{GWD}_u$ ,  $\text{GWD}_v$ , and  $\text{GWD}_T$  forcing as 3D Gaussian distribution have been chosen as  $-13 \text{ ms}^{-1}\text{day}^{-1}$ ,  $-0.13 \text{ ms}^{-1}\text{day}^{-1}$ , and  $0.065 \text{ Kday}^{-1}$  (see Tab. 1), respectively. The 3D Gaussian distribution for the H3 GW hotspot can be seen in the right panel of Fig. 2. In this paper we mainly concentrate on the 3D GW hotspots shaped like a box, when we analyse the effects on the middle atmosphere dynamics. For comparison regarding the shape of the artificial GW forcing, we just focus on the H3 GW hotspot with Gaussian smoothed boundaries, when we discuss the SPW modulation in section 3.2, which may be affected by the GW hotspots with sharp boundaries.

By comparing the size of the GW hotspots it is obvious (can be seen in Fig. 5) that the area of the enhanced GW drag, which scales with the cosine of the latitude, is decreasing for increasing latitude. But scaling the GW drag with latitude would lead to a much larger zonal mean GW drag at high latitudes and resulting changes in the circulation. Also the horizontal winds, which are affected by resulting nonlinear interactions are scaled in the model equations. In the current approach we are conserving the ratio of enhanced and unchanged GW drag values within the respective latitudinal belt, which is more meaningful. Also, the horizontal wavelength of PWs is smaller the closer to the pole they are, so that the ratio of the width of the GW forcing and the horizontal wavelength of the PWs remains the same for the respective latitudinal belt. In the following we will show that the spatial shape as well as the spatial size of the local GW forcing is not the most decisive factor, when we are comparing the 3D Gaussian distribution with the 3D GW forcing shaped as a box. Thus, GW hotspots having all the same size may lead to comparable results.

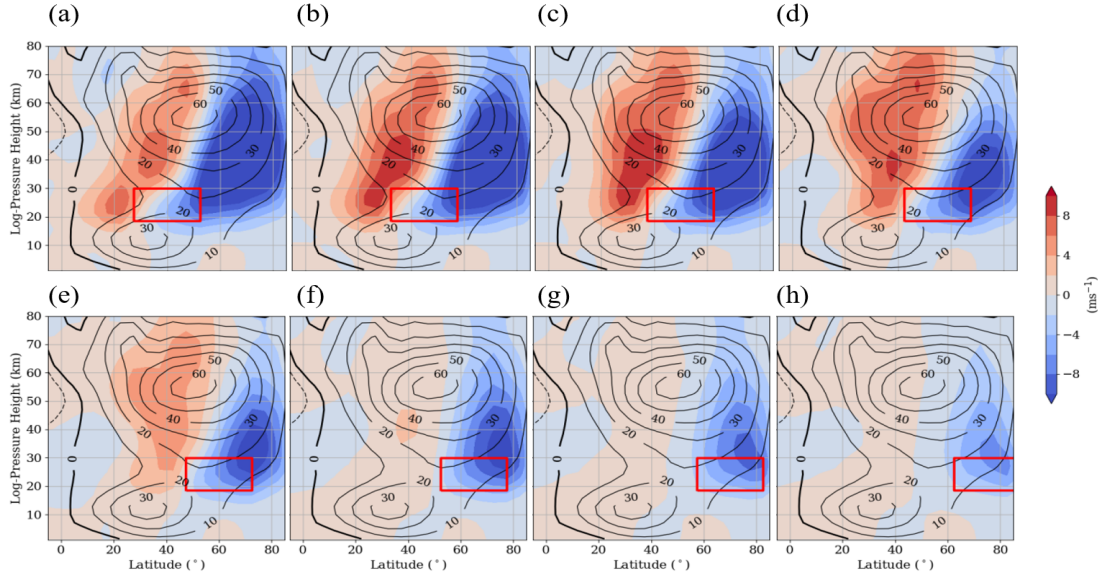


### 3 Results

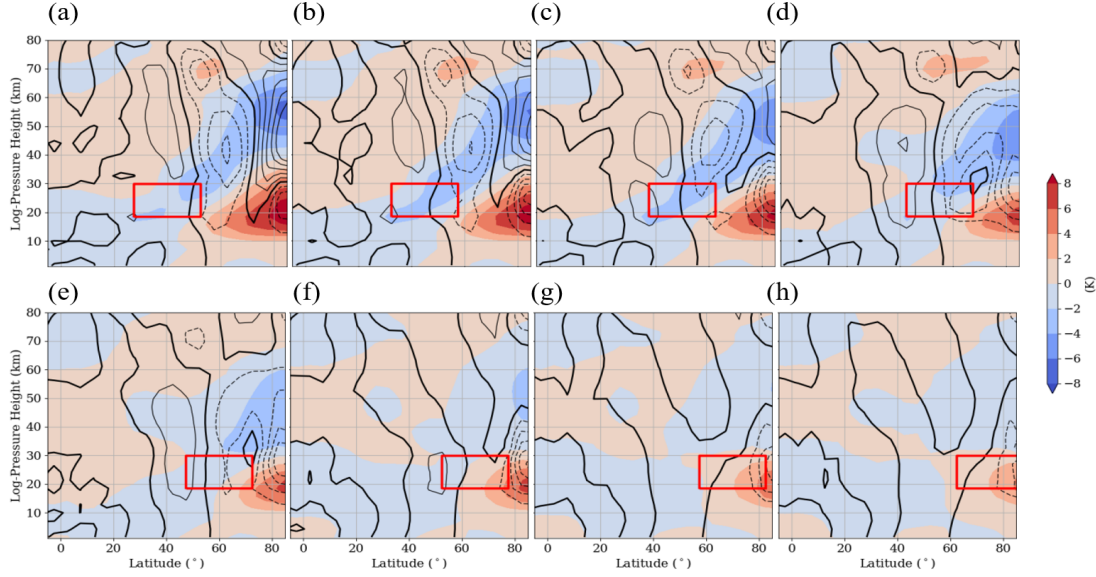
#### 3.1 Hotspot effect on the background circulation

Fig. 3 shows the zonal mean zonal wind difference between each GW hotspot simulation H1-H8 (a-h) and the Ref simulation in colour and the zonal mean zonal wind of the Ref simulation as contour lines in a latitude-height plot. The position of each GW hotspot is illustrated by the red box. All experiments (H1-H8) show negative zonal wind differences with a maximum wind decrease of more than  $-10 \text{ ms}^{-1}$  in the polar region. Positive differences can be observed equatorward from an imaginary line connecting the subtropical and polar night jet centers, with a maximum difference of 8 to  $10 \text{ ms}^{-1}$ . These zonal wind anomalies are consistent with a polar vortex that is shifted towards lower latitudes, and the wind reversal in the mesosphere is shifted upwards at lower latitudes. The strongest decrease of the zonal mean zonal wind in the polar region wind can be observed in the H1 simulation (a) and the strongest increase of zonal mean zonal wind at lower latitudes can be observed in the H3 simulation (c), the latter is corresponding to the observed Asian GW hotspot. For GW hotspots with a southern edge north of  $50^\circ\text{N}$  the polar vortex is only slightly displaced towards lower latitudes. Thus, the effect of GW hotspots at higher latitudes is less strong.

Fig. 4 is arranged in the same manner as Fig. 3 but shows the temperature difference in colour and the vertical wind difference as contour lines. As expected, the temperature effect scales with the (both zonal and vertical) wind differences, so that the H1-H3 simulations in Fig. 4(a-c) show the strongest temperature anomalies, and these are once more decreasing in magnitude for northward shifted GW hotspots.



**Figure 3.** Zonal mean zonal wind difference between the H1-H8 (a-h) and the reference simulation (H1-8 - Ref). Color indicates the difference between the simulations and the contour lines show the zonal mean zonal wind of the reference simulation. Figures represent the last 30 days of the simulations. The position of each GW hotspot is represented by a red box.

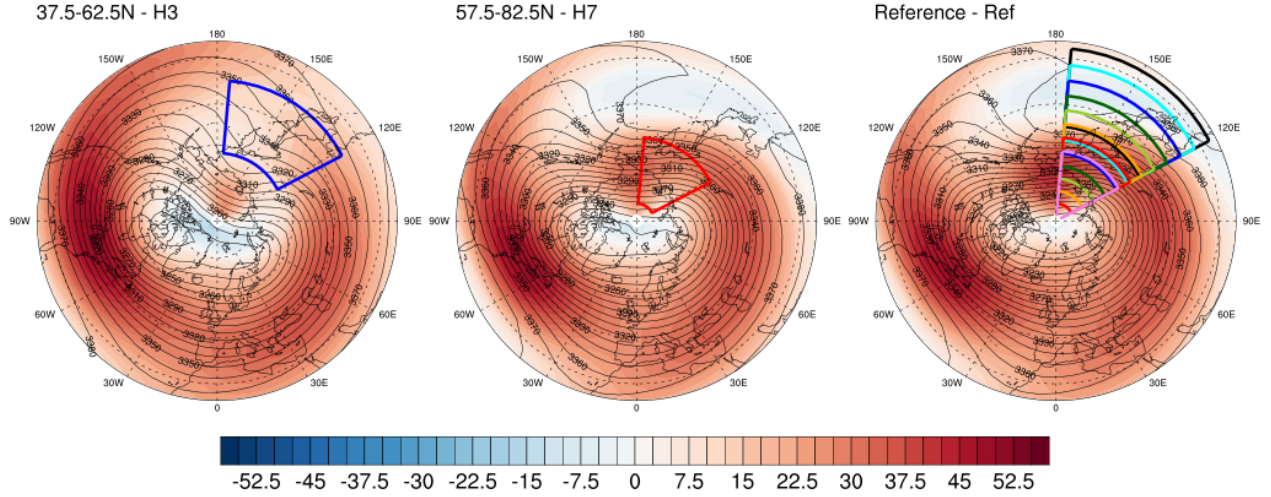


**Figure 4.** Zonal mean temperature and vertical wind difference between the H1-H8 (a-h) and the reference simulation (H1-8 – Ref). Color indicates the temperature difference between the simulations and the contour lines show the vertical wind difference starting with  $\pm 0.0005 \text{ ms}^{-1}$  up to  $\pm 0.0025 \text{ ms}^{-1}$  with increments of  $0.0005 \text{ ms}^{-1}$  and a thicker zero line. Negative (positive) values are shown by dashed (solid lines). Figures represent the last 30 days of the simulations. The position of each GW hotspot is represented by a red box.

Between  $60^\circ\text{N}$  and  $90^\circ\text{N}$ , the GW hotspot leads to a temperature increase at altitudes up to 30-35 km, but to a decrease above. The zonal mean vertical wind difference shows generally negative anomalies between 15 km and 30 km at higher latitudes, which indicates a stronger downward movement connected with an adiabatic warming in the lower part of the polar stratosphere. Above 35-40 km, we observe a positive vertical wind anomaly for the H1-H3 simulations, i.e. the downward movement is reduced and is leading to an adiabatic cooling anomaly. For most of the simulations, the negative anomaly in the lower part of the stratosphere is stronger than the positive anomaly above 40 km, which goes with the distribution of the temperature anomalies. In case of the H4 and H5 (Fig. 4e, f) simulations the vertical wind anomalies do not fit to the temperature anomalies. We observe an increased downward movement in a region, where the temperature is weakly decreasing.

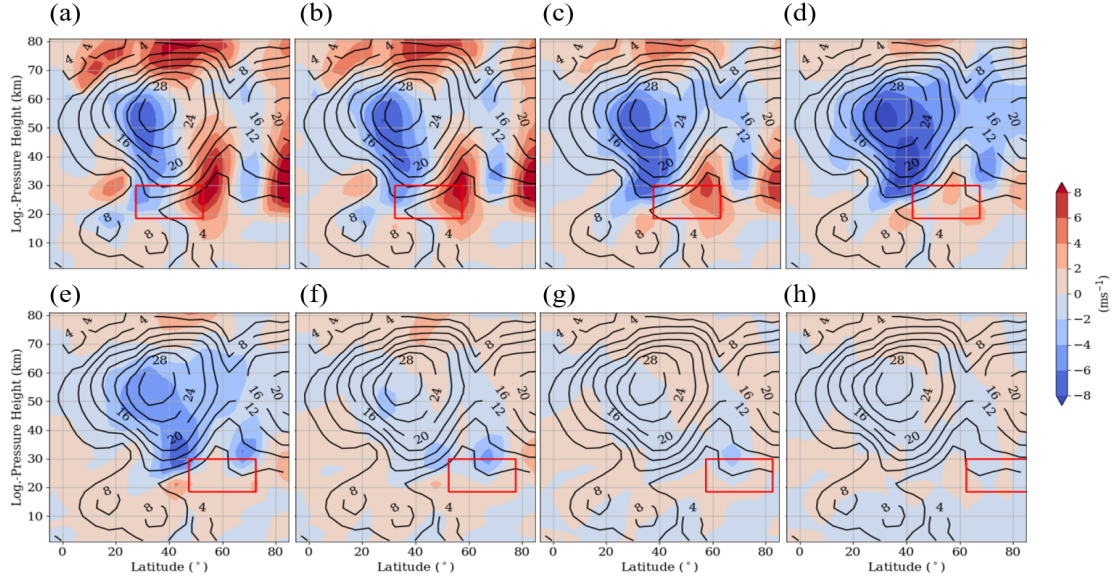
### 3.2 Influence on the polar vortex and anomalous SPWs

From previous publications it is already known that a warming (cooling) of the high-latitude stratosphere (mesosphere) and related changes in the dynamics are generally connected with PW activity. This leads us to the hypothesis that the main GWD enhancement effect is through SPW modulation, and this will be investigated in this subsection. In Fig. 5 we show for the H3 (left), H7 (middle) and Ref simulation (right panel) the geopotential height as contour lines and the zonal wind in colour coding as polar plot at 35 km, i.e., 5 km above the region of GW forcing. The panels represent the last 30 days of analysis. The position of each GW hotspot is illustrated by the boxes (H1 (black) - H8 (violet)) in the right panel of Fig. 5. The polar vortex of the Ref simulation is stable (not displaced or splitted) and located near the North Pole (right panel of Fig. 5).



**Figure 5.** Zonal wind ( $\text{ms}^{-1}$ ) in colour and geopotential height (gpdam) as contour lines northward of  $25^\circ\text{N}$  at 35 km for the H3 (left panel), H7 (middle panel) and the Ref simulation (right panel) representing the last 30 days of analysis. The boxes illustrate the position of each GW hotspot (H1 (in black) - H8 (in violet)). The blue (left and right panel) (red (middle and right panel)) box refers to the H3 (H7) simulation.

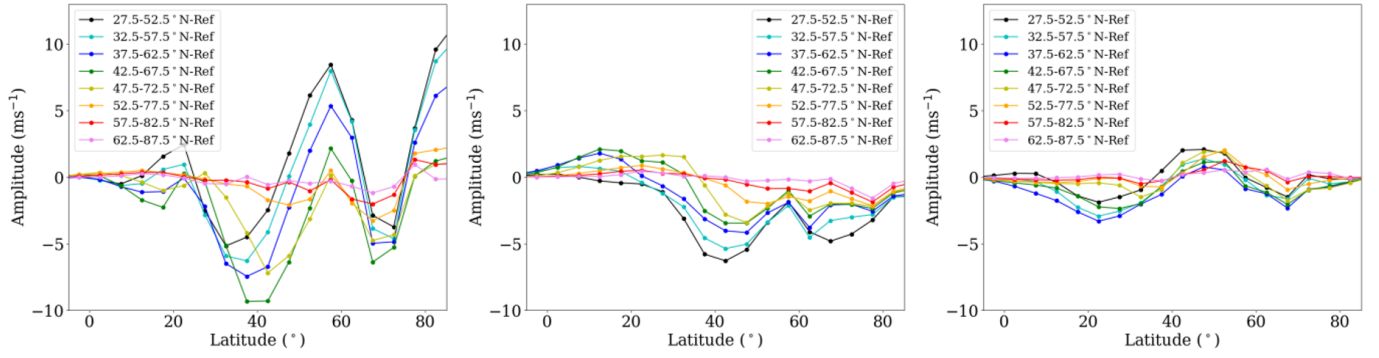
Between  $30$  and  $55^\circ\text{N}$  the zonal wind of the Ref simulation is easterly in one part of the EA/NP region due to the Aleutian high (AH). This means that the GW forcing, which is normally acting against the westerly zonal mean zonal wind, is locally strengthening the zonal wind. Between  $55^\circ\text{N}$  and  $90^\circ\text{N}$  there is a strong westerly wind between East Asia and Alaska, thus, the GW forcing there is acting locally against the zonal wind. Most of the GW hotspots (H1-H5) are located in the transition zone between easterlies and westerlies, whereby the south-eastern (north-western) part of each GW hotspot is located within the region of the easterlies (westerlies). The part of the GW forcing which is located in the easterlies (westerlies) is decreasing (increasing) for the more northward shifted GW hotspots (right panel of Fig. 5). In the H3 simulation, half of the GW hotspot is located in the easterlies and the other half in the westerlies. The state of the polar vortex of the H3 simulation is presented in the left panel of Fig. 5. The AH completely disappeared (no easterly wind anymore) and the center of the polar vortex is shifted towards Canada/Greenland. It is comma-like formed and slightly weaker and broader than the polar vortex of the Ref simulation. Thus, the H3 GW forcing has a destructive effect on the vortex circulation. This is in accordance with the results of the zonal wind differences (H3 - Ref) in Fig. 3(c) showing the displacement of the polar vortex edge to lower latitudes. The H7 GW hotspot (middle panel of Fig. 5), which is completely located in the westerlies, the polar vortex is less disturbed by the GW forcing and remains nearly at the same position as in the Ref simulation. The non-zonal part of the zonal wind field which is mainly dominated by the SPW 1 will interact with the local zonal wind anomaly induced by the localized GW forcing. Since this zonal wind anomaly is localized in longitude, it may be decomposed into a spectrum of harmonics and can be assumed to be an additional wave interacting with the original zonal wind SPW 1. To examine this interaction between the original SPW 1 in the model and the one induced by the local GW forcing the difference of the SPW 1 zonal wind amplitude between each GW hotspot simulation H1-H8 and the Ref simulation is shown in Fig. 6.



**Figure 6.** Zonal mean SPW 1 amplitude extracted from the zonal wind as difference between the H1-H8 (a-h) and the reference simulation. Color indicates the difference between the simulations and the contour lines show the zonal mean SPW 1 amplitude of the reference simulation. Figures represent the last 30 days of the simulations. The position of each GW hotspot is represented by the red box.

The position of each GW hotspot is illustrated by the red boxes. In the Ref simulation the SPW 1 amplitude is maximizing at about 55 km between 30°N and 40°N with more than  $28 \text{ ms}^{-1}$  and also in the polar stratosphere with about  $20 \text{ ms}^{-1}$ . In case of the H1-H4 simulations in Fig. 6(a-d) the zonal wind SPW 1 amplitude differences are positive (negative) on the northern (southern) flank of the respective GW hotspot up to an altitude of about 60 km. The negative (positive) SPW 1 amplitude anomaly is increasing (decreasing) for more northward GW hotspots. The strongest increase (decrease) of SPW 1 amplitude can be observed in the H1 (H4) simulation with more than  $8$  ( $-8 \text{ ms}^{-1}$ ). By comparing the positive and negative SPW 1 amplitude anomalies of the H1-H4 simulations it can be seen that the positive anomaly is less pronounced, while the negative anomaly is more prevalent in the whole NH, particularly around the stratopause. The decreasing SPW 1 amplitude indicates that less SPWs 1 are propagating into the middle atmosphere. Due to the decreasing SPW 1 activity at lower latitudes, less SPWs 1 are breaking in this region, i.e. the zonal mean zonal wind is less decelerated as it is shown in Fig. 3. Non-locally, however, a localized destructive (constructive) superposition of the original SPWs 1 within the model and the one of the GW forcing may decrease (increase) the SPW 1 amplitude at other heights/latitudes due to changes of PW propagation. This effect can be observed around 55°N, where we observe an enhanced SPW 1 amplitude. It is strongest for the H1 GW hotspot and is decreasing for more northward located GW hotspots. The suppressed upward propagation of SPWs 1 is leading to the increasing SPW 1 amplitude in this area. This positive SPW 1 amplitude anomaly is corresponding to the decelerated zonal mean zonal wind in Fig.3. This leads to the assumption that the GW forcing may locally in- or decrease the SPW 1 amplitude but prevents the SPWs from propagating upwards into higher altitudes so that the SPW 1 amplitude is mainly decreasing in the stratosphere/mesosphere. Thus, the local GW forcing has a destructive effect on the circulation in the middle atmosphere.

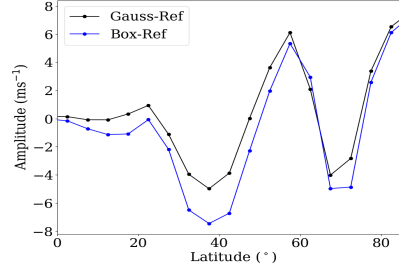




**Figure 7.** Zonal mean SPW 1 (left panel), SPW 2 (middle panel) and SPW 3 (right panel) amplitude as difference between the H1-H8 and the Ref simulation at 35 km for the last 30 days of the simulations extracted from the zonal wind.

We will verify this in section 3.3 below by analysing the Eliassen-Palm flux. Owing to the suppression of SPW 1 propagation at midlatitudes, the SPWs may increasingly propagate via the polar region, which may explain the increased SPW 1 amplitude in the polar stratosphere northward of 75°N. Another positive SPW 1 amplitude anomaly can be observed in the midlatitudinal mesosphere above 60 km which may be induced by local instabilities generating new SPWs 1. Both of these positive SPW 1 amplitude anomalies are strongest for the H1 simulation and are once more decreasing for northward shifted GW hotspots. The SPW 1 amplitude anomalies for the four northernmost GW hotspots H5-H8 in Fig. 6(e-h) are small in comparison to the four southernmost GW hotspot simulations, which corresponds to the observations in section 3.1. Only for the H5 simulation (Fig. 6(e)) the SPW 1 activity is also strongly reduced at lower latitudes above 30 km like in the H1-H4 simulations.

To analyse in how far the GW forcing is locally affecting the SPWs of wavenumber 2 and 3, we compared the SPW 1 (left), 2 (middle), and 3 (right panel) amplitude anomalies at 35 km northward of 0°N/S (Fig. 7). The colours are the same as the colors of the hotspots in Fig. 5. As already discussed in Fig. 6, the SPW 1 amplitude locally increases at midlatitudes and in the polar region with about 10 ms<sup>-1</sup> in maximum whereby the maxima are decreasing for northward displaced GW hotspots. The negative anomaly, which is mainly dominating the middle atmosphere is located between 30°N and 40°N as well as at 70°N with more than -10 ms<sup>-1</sup> in minimum. The SPW 2 amplitude it can be seen that the SPW 2 activity is weakened/reduced northward of 30°N by about -6 ms<sup>-1</sup> in minimum. The largest decrease can be observed for the southernmost GW hotspot, and the decrease is getting smaller for northward displaced GW hotspots. Only at lower latitudes the SPW 2 amplitude is slightly increasing for those simulations. This is the case for the H2, H3, H4 and H5 simulation. The SPW 2 anomaly is negative (positive) in those regions, where the SPW 1 anomaly is positive (negative). This leads to the assumption that just one of both SPWs (SPW 1 and SPW 2) can be the dominating one. By comparing the latitudinal distribution of the SPW 1 and 2 amplitude anomalies northward of 30°N it can be seen that they are similar when we neglect the scales. Both show a decrease in amplitude around 40°N and 70°N and an increase in the midlatitudes and in the polar region. In comparison, the SPW 3 amplitude anomaly distribution (Fig. 7 right) is slightly different because the SPW 3 amplitude is decreasing at 20°N (not at 40°N as for the SPW 1 and 2 anomalies).



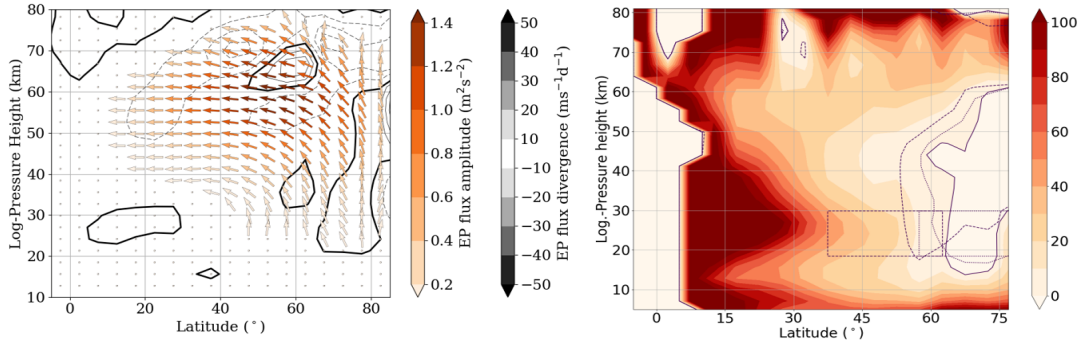
**Figure 8.** SPW 1 amplitude as difference between the H3 and the reference simulation (blue line) and the Gauss distribution and the reference simulation (black line) at 35 km for the last 30 days extracted from the zonal wind.

But, as for the SPW 1 and 2 anomalies, an increase of the SPW 3 amplitude induced by the local GW hotspots can be observed in the midlatitudes with about  $2 \text{ ms}^{-1}$  in maximum. The largest increase in SPW 3 amplitude can be seen in the H1 simulation (southernmost GW hotspot) and the largest decrease in the observed H3 simulation (observed Asian GW hotspot). The suppression of SPWs, which is induced by the local GW forcing, might be also an effect partly induced by the shape of the GW hotspot leading to a sharp transition zone between the unchanged and enhanced GW drag values. To prove if the shape of the GW hotspot is partly leading to a suppression of SPWs, Fig. 8 shows the H3 amplitude anomaly from the left panel of Fig. 7 together with the corresponding Gauss simulation described in section 2. The latitudinal distribution of the SPW 1 amplitude difference is still the same showing the two local maxima at midlatitudes and in the polar region and the two minima at  $40^\circ\text{N}$  and  $70^\circ\text{N}$  but these two minima have changed from  $-8 \text{ ms}^{-1}$  for the three dimensional box to  $-4 \text{ ms}^{-1}$  for the Gaussian distribution. Also the maxima have changed from about  $4 \text{ ms}^{-1}$  to more than  $5 \text{ ms}^{-1}$ . Due to the stronger maximum GW drag in the Gaussian distribution the SPW 1 excitation is strengthened, which is leading to the larger SPW 1 amplitudes at midlatitudes. The smoothly decreasing GW drag forcing towards lower and higher latitudes only slightly reduces the suppression of SPW 1 around  $40^\circ\text{N}$  and  $70^\circ\text{N}$ . Also, the mean wind and temperatures are only weakly affected, if we replace the box-like forcing by one with a Gaussian shape (not shown here). Thus, the GW hotspot itself is leading to essential changes in the dynamics suppressing the SPW propagation and decreasing the SPW 1 activity in the middle atmosphere.

To sum up the influence on the polar vortex, in section 3.1 we have observed a slight warming of the lower stratosphere and the decreasing west wind at middle to high latitudes, which indicates a weakening of the polar vortex as a consequence of the GW drag enhancement. Thus, the stability of the polar vortex is not only depending on the PW activity but rather on the interplay or nonlinear interaction of GW and PW forcings. The anomalous SPW forcing and suppression of SPW propagation show that the GW drag can play an important role for the preconditioning of the polar vortex (see next section).

### 3.3 Propagation conditions for SPWs

To see in how far the SPW propagation is affected by the local GW forcing the EP fluxes and their divergence for SPW 1 were calculated. The results of the Ref simulation are presented in Fig. 9 (left panel). The arrows show the direction of propagation, the colour of the arrows represents the strength of the EP flux not normalized by the density, and the grey areas as well as the

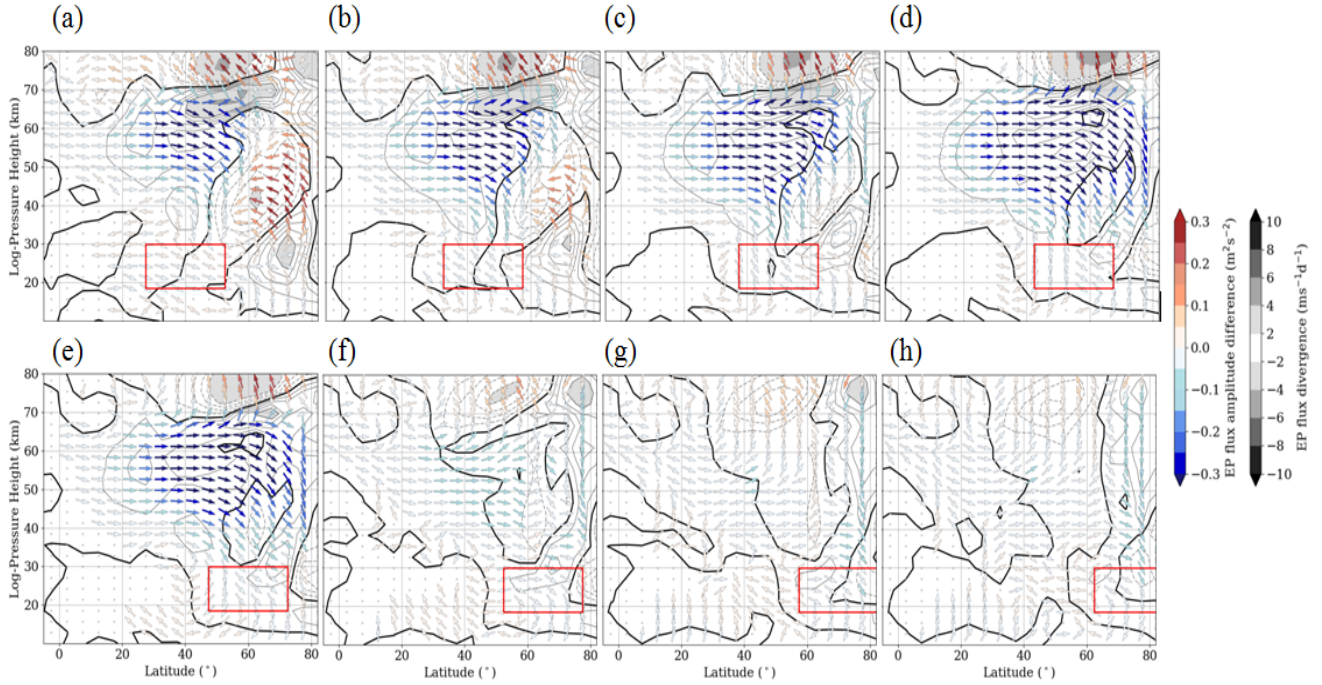


**Figure 9.** Zonal mean EP flux of SPW 1 of the Ref simulation (left panel). Contour lines show EP flux divergence, dashed lines denote negative EP flux divergence. Refractive index for SPW 1 for the Ref simulation (right panel) with a thicker zero line. The position of the H3 (H7) GW hotspot and the respective zero line is represented by the dashed (dotted) line colored in violet. Both Figs. represent the last 30 days of the simulation.

grey contour lines represent the EP divergence showing in which direction the zonal mean flow is accelerated. A negative (positive) EP divergence is illustrated by the dashed (solid) lines. The arrows were replaced by a dot when the amplitude is smaller than 1% of the maximum EP flux amplitude. The waves are mainly developing in the mid- and higher latitudes and from there they are propagating mainly towards the equatorial stratosphere/stratopause and, to a much lesser degree to the polar stratosphere. That the waves are really propagating upwards can be seen by means of the increasing amplitudes of the EP fluxes. The maximum EP flux amplitudes of more than  $1.4 \text{ m}^2\text{s}^{-2}$  are reached between  $50^\circ\text{N}$  and  $60^\circ\text{N}$  at an altitude of about 60 km, which corresponds to the height of the SPW 1 amplitude maximum in Fig. 1(f) of the Ref simulation.

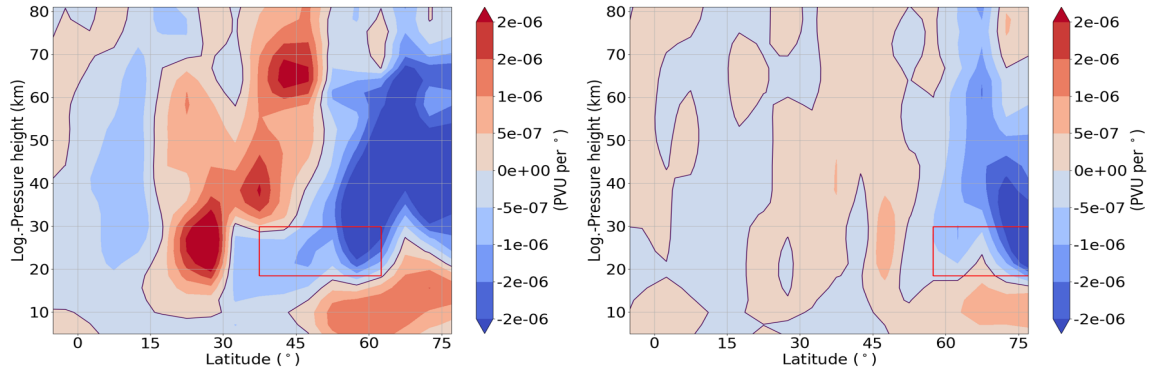
In Fig. 10 the difference of the EP flux and its divergence between the H1-H8 and the Ref simulation is shown. The position of each GW hotspot is again illustrated by the red boxes. In the H1 and H2 simulations (Fig. 10(a-b)) more SPWs 1 are propagating into the polar stratosphere. These SPWs 1 are partly coming from the midlatitudes but most of them are directly generated in the polar region, where we observe a source of SPWs 1 (enhanced positive EP divergence at  $70^\circ\text{N}$  between 20 and 30 km). This positive EP divergence anomaly corresponds to the increased SPW 1 amplitude in the polar region in Fig. 7. Above this positive EP divergence anomaly an enhanced negative EP divergence is seen (from the northern flank of the GW hotspot up to 60 km tilted towards the North with increasing height), which means that the SPWs 1, which are propagating via the Arctic stratosphere, are breaking in this region. This leads to the deceleration of the zonal mean zonal wind in the middle and higher latitudes which was already discussed in Fig. 3. The negative EP divergence anomaly can also be seen in the H3-H5 simulations (Fig. 10(c-e)). This is the reason why the polar vortex is mainly disturbed by these GW hotspots (H1-H5). The negative EP divergence is strongest for the H1 simulation, which also exhibits the strongest increase of SPW 1 amplitude in the polar region. Furthermore, the H1-H5 simulations show a strong decrease of the EP flux amplitude (blue arrows) between 40 km and 70 km and  $20^\circ\text{N}$  to  $80^\circ\text{N}$ , which means that less SPWs 1 are propagating into the middle atmosphere. As a consequence, less SPWs 1 are breaking in this region leading to a positive EP divergence anomaly.





**Figure 10.** Zonal mean EP flux (arrows) and divergence (isolines and shaded areas, dashed lines show negative values) of SPW 1. Shown is the difference between all H1-8 (a-h) simulations and the reference simulation (H1-8 – Ref) representing the last 30 days of the simulations.

This result corresponds to the decreasing SPW 1 amplitude (Fig. 7) and the increasing zonal mean zonal wind (Fig. 3) at lower latitudes. The effect is strongest for the H4 simulation, which also shows the strongest decrease in SPW 1 amplitudes. Between 40°N and 70°N around 70 km we observe another source of SPWs 1, which propagate into the mesosphere, where these waves are breaking (strongly negative EP divergence above the positive EP divergence) due to the reversed wind conditions. The mesospheric EP flux in the H1-H5 simulations corresponds to the observed enhanced SPW 1 amplitude in the mesosphere in Fig. 7. Referring to the enhanced SPW 1 around 55°N of the GW hotspots no enhanced EP flux can be observed in the respective region. But the arrows of the EP flux anomalies are pointing towards this area of enhanced SPW 1 amplitude. In the H6-H8 GW hotspot simulations (Fig. 10(f-h)) no large differences in EP flux and divergence occur, which corresponds to the small SPW 1 amplitude and the zonal mean zonal wind differences in Fig.7 and Fig.3. To explain why SPWs 1 do not propagate at higher latitudes also the refractive index (Matsuno, 1971; Andrews et al., 1987), multiplied by the square of the Earth's radius  $a^2$ , is shown in Fig. 9 (right panel). The refractive index highly depends on the meridional potential vorticity gradient ( $q_y$ ) and on the zonal mean zonal wind conditions (Li et al., 2007). White regions in Fig. 9 (right panel) indicate a negative refractive index, which means that the waves cannot propagate in these regions. In the reddish regions wave propagation is possible. Due to the predominating westerly wind in the northern hemisphere the refractive index is mostly positive, so that SPWs 1 are able to propagate predominantly upward and towards the equator. Towards the midlatitudes and the polar region the refractive index is decreasing because of the increasing zonal mean zonal wind.



**Figure 11.** Meridional potential vorticity gradient difference between the H3 (left panel), H7 (right panel) and the reference simulation representing the last 30 days of the simulations. The positions of the H3 and H7 GW hotspot is represented by the red boxes.

The polar region (northward of  $60^\circ\text{N}$ ) is the only region with a negative refractive index in the northern hemisphere. This is because the polar vortex is a strong closed system, which repels most of the waves. To see in how far the refractive index is changing after implementing the GW forcing the position of the H3 (dashed line colored in violet) and the H7 (dotted line colored in violet) GW hotspot and the respective zero line of the refractive index were added to Fig. 9 (right panel). In the H3 simulation the zero line is higher in the polar region than the one in the Ref simulation. Thus, the refractive index is increasing (becomes more positive) in the polar region below 30 km, which corresponds to the enhanced SPW 1 propagation and SPW 1 amplitude in the same region. The zero line of the H7 simulation is nearly at the same height as the one of the Ref simulation so that we do not observe huge changes in the Arctic. While the zero line of the Ref simulation is limited to the regions northward of  $60^\circ\text{N}$ , the zero line of the H3 (H7) simulation is located around  $50^\circ\text{N}$  ( $57^\circ\text{N}$ ). Based on the EP flux distribution of the Ref simulation in Fig. 9 (left panel) we already know that the SPWs 1 are mainly propagating from the midlatitudes (between  $50$  and  $60^\circ\text{N}$ ) into the middle atmosphere. Due to the negative refractive index in this region the SPWs 1 in the H3 and H7 simulations are not able anymore to propagate upwards so that the SPW 1 EP flux and amplitude are decreasing. Thus, the major branch of SPW 1 propagation is interrupted by the local GW forcing. To check if there are local instabilities leading to the SPW 1 sources in the polar region and in the lower mesosphere, the  $q_y$  differences between the H3 (H7) and the Ref simulation are shown in Fig. 11. The  $q_y$  is given in potential vorticity units (PVU) per degree. The positions of the H3 and H7 GW hotspot are illustrated by the red boxes. Due to the increasing (decreasing) zonal mean zonal wind at lower (higher) latitudes, the  $q_y$ , which is normally increasing towards higher latitudes, is reversed northward of  $30^\circ\text{N}$ . We observe a negative  $q_y$  anomaly which is tilted towards the North with increasing height. Northward of  $45^\circ\text{N}$  up to 20 km the  $q_y$  anomaly reverses again and becomes positive. These local reversals of the  $q_y$ , which are a necessary condition for baroclinic instability (Charney and Stern, 1962), can lead to the SPW 1 sources and positive EP divergences in the respective regions.

#### 4 Conclusions

The sensitivity study regarding the effect of local GW hotspots in the stratosphere from lower to higher latitudes in a specific longitude range between 120°E and 170°E shows that GW hotspots southward of 50°N are leading to a negative refractive index at midlatitudes which prevents the SPWs from propagating upwards. Thus, less SPW 1 are breaking in the middle atmosphere corresponding to the decreasing SPW 1 amplitude at lower latitudes connected with an increasing zonal mean zonal wind. Thus, the polar vortex is shifted towards lower latitudes but remains really strong (Baldwin and Holton, 1988) and leads additionally to a suppression of SPWs according to the Charney-Drazin criterion (Charney and Drazin, 1961). The displacement of the polar vortex induced by breaking SPWs 1 causes an increase of the refractive index in the polar stratosphere (Karami et al., 2016) so that the SPWs 1 originating at midlatitudes are partly propagating via the polar region into the middle atmosphere. Apart from these SPWs 1, additional SPWs 1 are propagating upwards which are directly generated in the Arctic owing to local baroclinic instability. One indication is the reversal of the  $q_y$  (Charney and Stern, 1962; Garcia, 1991). For this reason we observe an enhanced EP flux and thus, an enhanced SPW 1 amplitude in the polar region. These SPWs 1 are breaking around 50 km between 50 and 80°N and lead to an enhanced negative EP divergence connected with a decreasing zonal mean zonal wind at higher latitudes. In the lower mesosphere between 40 and 70°N there is a second source of SPW 1 (positive EP divergence) developed as well due to local baroclinic instabilities (reversal of the  $q_y$ ) (Smith, 2003; Lieberman et al., 2013; Matthias and Ern, 2018). As a consequence, the EP flux and the SPW 1 amplitude are enhanced between 70 and 80 km, right above the positive EP divergence anomaly. Based on the SPW 1 amplitude extracted from the zonal GW drag (not shown here) it was clear that each of the GW hotspots leads to a forcing of SPWs 1 but in some regions northward of 50°N this forcing is ineffective because the waves cannot propagate or are eliminated by destructive interference. The refractive index, which is highly depending on the zonal mean zonal wind conditions, shows negative values in the polar region for the Ref simulation. So, if we implement a GW forcing directly in this region it has no impact on the middle atmosphere because SPWs cannot propagate. If we provoke a preconditioning of the polar vortex by first implementing e.g. the H1 GW hotspot and then adding one of the H6-H8 GW hotspots then the GW hotspots near the polar region would have a larger impact on the dynamics on the middle atmosphere.

Based on these results of the sensitivity study we have seen that a local GW forcing can lead to a weakening (warming of the lower stratosphere and slight displacement) of the polar vortex at high latitudes, which is highly depending on the strength (Šácha et al., 2016) as well as on the zonal distribution of the forcing (this study). Usually, it is assumed that the preconditioning of the polar vortex is mainly driven by enhanced PW activity (Labitzke, 1981). But there are also several indications based on satellite observations (Ern et al., 2016) and reanalysis data (Albers and Birner, 2014) showing that the GW drag and the absolute GW momentum flux is enhanced (reduced) in the stratosphere right before (after) SSWs. Albers and Birner (2014) have analyzed the total wave forcing from Japanese Meteorological Agency and Central Research Institute of Electrical Power Industry 25-year Reanalysis (JRA-25) data before SSW events and found that up to 70% of the total drag is induced by orographic GWs. Ern et al. (2016) directly derived the GW drag and absolute momentum fluxes from HIRDLS and SABER temperatures and found out that both parameters are enhanced before and around the central day of a SSW (strong polar jet)

and reduced, when the zonal wind is weak (after SSW). Because we kept the GW drag forcing constant during the whole experiments we cannot evaluate nonlinear effects, which would possibly reduce the GW drag connected with the displacement of the polar vortex. Furthermore, we have a fixed GW source distribution so that no additional GWs are generated owing to changes in the tropospheric circulation. But on the basis of the zonal and meridional GW flux, which is changing according to the propagation conditions, we have analyzed the absolute horizontal GW momentum flux (not shown here). In this case, we observe as well a reduction of the GW flux, when the zonal mean zonal wind is decreasing at high latitudes. However, this effect is not very pronounced in our experiments because the zonal mean zonal wind differences are much smaller than during a real SSW event. We only observe zonal mean zonal wind differences of about  $-10 \text{ ms}^{-1}$ , which do not lead to a reversal of the zonal mean zonal wind, and thus, to significant background changes, which may strongly influence the GW propagation.

Another interesting aspect is that different shapes of the local GW forcing will not have strong effects on the circulation. Despite of Gaussian-smoothed boundaries just negligible changes can be observed in the dynamics and SPW development, which are mainly due to the varying GW drag in the three dimensional Gaussian distribution leading to larger (smaller) effects when the Gaussian distribution is maximizing (minimizing).

Comparing the positions of these simulated GW hotspots with measurements (e.g. Hoffmann et al., 2013) it is clear that at least some of the latitudinally shifted GW hotspots are not very realistic so that our experiments should only be considered as a qualitative sensitivity study. Regarding orography there are no obvious sources, when we displace the GW hotspot latitudinally. But indeed, some of these GW hotspots can be purely hypothetical connected to jet exit regions. To make it more realistic, the next step will be to analyse the effect of a longitudinally shifted hotspot (fixed latitude range between  $30^\circ\text{N}$  and  $60^\circ\text{N}$ ), because observations and GCM experiments have shown their existence (listed in the introduction). In this latitude range GW hotspots like the Himalayan region, the Alps or the Rocky mountains are part of the experiments. Also the interaction of two or more GW hotspots is part of our interests and will provide more insights into the effect of a localized GW forcing, which may be also important for the development of new GW parametrizations.

*Code availability.* MUAM model code is available from the corresponding author upon request.

*Author contributions.* N. Samtleben performed the MUAM model simulations and drafted the first version of the manuscript. P. Šácha provided GW potential energy distributions. P. Šácha, P. Pišoft, A. Kuchař and C. Jacobi actively contributed to the discussions and the writing of the paper.

*Competing interests.* C. Jacobi is one of the Editors-in-Chief of *Annales Geophysicae*. The authors declare that there are no conflicts of interest.

*Acknowledgements.* This study has been supported by Deutsche Forschungsgemeinschaft (DFG) under grant JA836/32-1, and by GA CR under grant 16-01562J. ERA Interim reanalysis data have been provided by ECMWF through [apps.ecmwf.int/datasets/data/](https://apps.ecmwf.int/datasets/data/). Petr Pišoft and Aleš Kuchař were supported by GA CR under grant nos. 16-01562J and 18-01625S. Petr Šácha was supported by GA CR under grant nos. 16-01562J and 18-01625S and by Government of Spain under grant no. CGL2015-71575-P and in later stages of the manuscript preparation through a postdoctoral grant of the Xunta de Galicia ED481B 2018/103. The EPhysLab is supported through the European Regional Development Fund (ERDF).

## References

- Albers, J. R. and Birner, T.: Vortex Preconditioning due to Planetary and Gravity Waves prior to Sudden Stratospheric Warmings, *J. Atmos. Sci.*, 71, 4028–4054, <https://doi.org/10.1175/JAS-D-14-0026.1>, 2014.
- Alexander, P., Luna, D., Llamedo, P., and de la Torre, A.: A gravity wave study close to the Andes mountains in Patagonia and Antarctica  
5 with GPS radio occultation observations, *Ann. Geophys.*, 28, 587–595, <https://doi.org/10.5194/angeo-28-587-2010>, 2010.
- Andrews, D. G., Holton, J. R., and Leovy, C. B.: *Middle Atmosphere Dynamics*, <https://doi.org/ISBN: 0-12-058576-6>, 1987.
- Baldwin, M. P. and Holton, J. R.: Climatology of the stratospheric polar vortex and planetary wave breaking, *J. Atmos. Sci.*, 45, 1123–1142, [https://doi.org/10.1175/1520-0469\(1988\)045<1123:COTSPV>2.0.CO;2](https://doi.org/10.1175/1520-0469(1988)045<1123:COTSPV>2.0.CO;2), 1988.
- Charney, J. G. and Drazin, P. G.: Propagation of planetary-scale disturbances from the lower into the upper atmosphere, *J. Geophys. Res.*, 66,  
10 83–109, <https://doi.org/10.1029/JZ066i001p00083>, 19, 20, 41, 62, 63, 71, 1961.
- Charney, J. G. and Stern, M. E.: On the Stability of Internal Baroclinic Jets in a Rotating Atmosphere, *J. Atmos. Sci.*, 19, 159–172, [https://doi.org/10.1175/1520-0469\(1962\)019<0159:OTSOIB>2.0.CO;2](https://doi.org/10.1175/1520-0469(1962)019<0159:OTSOIB>2.0.CO;2), 1962.
- Constantino, L., Heinrich, P., Mze, N., and Hauchecorne, A.: Convective gravity wave propagation and breaking in the stratosphere: comparison between WRF model simulations and lidar data, *Ann. Geophys.*, 33, 1155–1171, <https://doi.org/10.5194/angeo-3331155-2015>,  
15 2015.
- Dee, D. P., Uppala, S. M., Simmons, A. J., Berrisford, P., Poli, P., Kobayashi, S., Andrae, U., Balmaseda, M. A., Balsamo, G., Bauer, P., Bechtold, P., Beljaars, A. C. M., van de Berg, L., Bidlot, J., Bormann, N., Delsol, C., Dragani, R., Fuentes, M., Geer, A. J., Haimberger, L., Healy, S. B., Hersbach, H., Hólm, E. V., Isaksen, I., Kallberg, P., Köhler, M., Matricardi, M., McNally, A. P., Monge-Sanz, B. M., Morcrette, J.-J., Park, B.-K., Peubey, C., de Rosnay, P., Tavolato, C., Thépaut, J.-N., and Vitart, F.: The ERA-Interim reanalysis: configuration  
20 and performance of the data assimilation system, *Quart. J. Roy. Meteor. Soc.*, 137, 553–597, <https://doi.org/10.1002/qj.828>, 2011.
- Douville, H.: Stratospheric polar vortex influence on Northern Hemisphere winter climate variability, *Geophys. Res. Lett.*, 36, L18 703, <https://doi.org/10.1029/2009GL039334>, 2009.
- Ern, M. and Preusse, P.: Gravity wave momentum flux spectra observed from satellite in the summertime subtropics: Implications for global modeling, *Geophys. Res. Lett.*, 39, <https://doi.org/10.1029/2012GL052659>, 2012.
- Ern, M., Preusse, P., Alexander, M. J., and Warner, C. D.: Absolute values of gravity wave momentum flux derived from satellite data, *J. Geophys. Res.*, 109, <https://doi.org/10.1029/2004JD004752>, 2004.
- Ern, M., Ploeger, F., Preusse, P., Gille, J. C., Gray, L. J., Kalisch, S., Mlynchak, M. G., III, J. M. R., and Riese, M.: Interaction of gravity waves with the QBO: A satellite perspective, *J. Geophys. Res. Atmos.*, 119, 2329–2355, <https://doi.org/10.1002/2013JD020731>, 2014.
- Ern, M., Trinh, Q. T., Kaufmann, M., Krisch, I., Preusse, P., Ungermann, J., Zhu, Y., Gille, J. C., Mlynchak, M. G., III, J. M. R., Schwartz,  
30 M. J., and Riese, M.: Satellite observations of middle atmosphere gravity wave absolute momentum flux and of its vertical gradient during recent stratospheric warmings, *Atmos. Chem. Phys.*, 16, 9983–10 019, <https://doi.org/10.5194/acp-16-9983-2016>, 2016.
- Fleming, E. L., Chandra, S., Barnett, J. J., and Corney, M.: Zonal mean temperature, pressure, zonal wind and geopotential height as functions of latitude, *Adv. Space Res.*, 10, 12, (12)11–(12)59, [https://doi.org/10.1016/02731177\(90\)90386-E](https://doi.org/10.1016/02731177(90)90386-E), 1988.
- Fomichev, V. I. and Shved, G. M.: Parameterization of the radiative flux divergence in the 9.6 $\mu$ m O<sub>3</sub> band, *J. Atmos. Terr. Phys.*, 47, 1037–  
35 1049, [https://doi.org/10.1016/0021-9169\(85\)90021-2](https://doi.org/10.1016/0021-9169(85)90021-2), 1985.
- Fomichev, V. I., Blanchet, J.-P., and Turner, D. S.: Matrix parameterization of the 15 mm CO<sub>2</sub> band cooling in the middle and upper atmosphere for variable CO<sub>2</sub> concentration, *J. Geophys. Res.*, 103, 11.505–11.528, <https://doi.org/10.1029/98JD00799>, 1998.

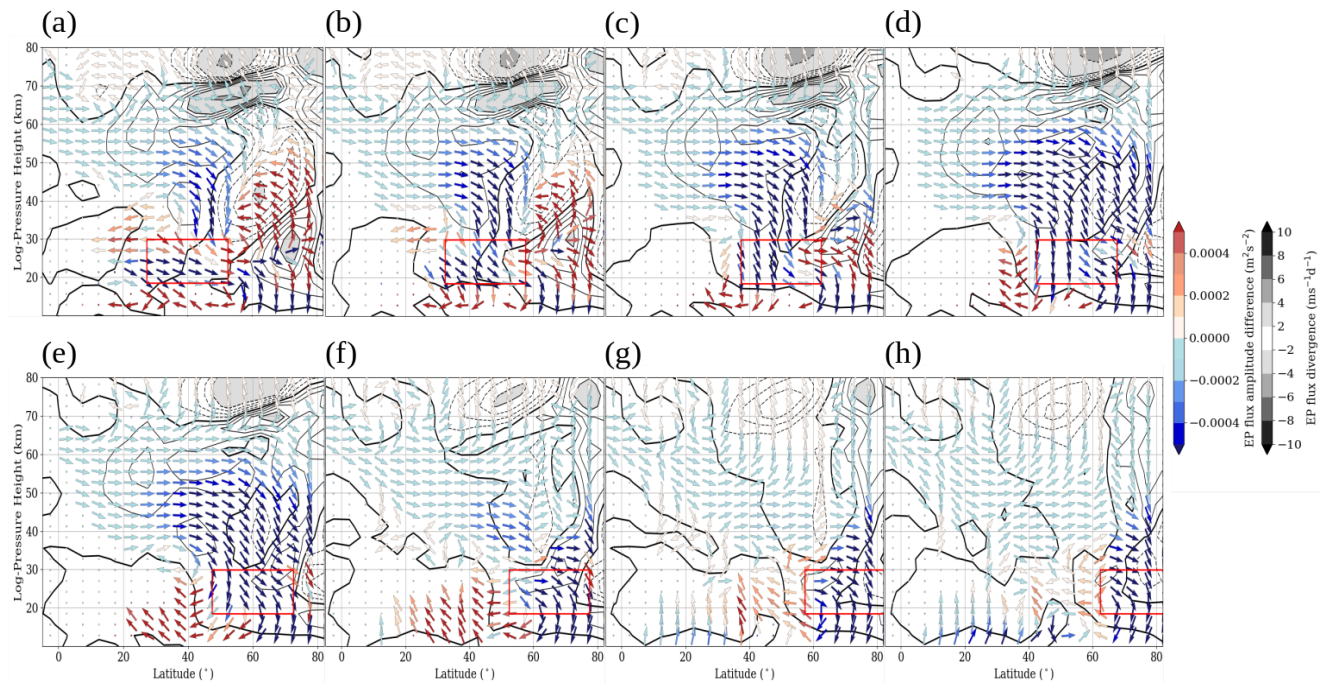
- Fritts, D. C. and Alexander, M. J.: Gravity wave dynamics and effects in the middle atmosphere, *Rev. Geophys.*, 41, 1003, <https://doi.org/10.1029/2001RG000106>, 2003.
- Fröhlich, K., Pogoreltsev, A., and Jacobi, C.: Numerical simulation of tides, Rossby and Kelvin waves with the COMMA-LIM model, *Adv. Space Res.*, 32, 863–868, [https://doi.org/10.1016/S0273-1177\(03\)00416-2](https://doi.org/10.1016/S0273-1177(03)00416-2), 2003a.
- 5 Fröhlich, K., Pogoreltsev, A., and Jacobi, C.: The 48-layer COMMA-LIM model, *Rep. Inst. Meteorol. Univ. Leipzig*, 30, 157–185, <https://doi.org/http://nbn-resolving.de/urn:nbn:de:bsz:15-qucosa-217766>, 2003b.
- Fröhlich, K., Schmidt, T., Ern, M., Preusse, P., de la Torre, A., Wickert, J., and Jacobi, C.: The global distribution of gravity wave energy in the lower stratosphere derived from GPS data and gravity wave modelling: attempt and challenges, *J. Atmos. Sol.-Terr. Phys.*, 69, 2238–2248, <https://doi.org/10.1016/j.jastp.2007.07.005>, 2007.
- 10 Garcia, R. R.: Parameterization of planetary wave breaking in the middle atmosphere, *J. Atmos. Sci.*, 48, 1405–1419, [https://doi.org/10.1175/1520-0469\(1991\)048<1405:POPWBI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1991)048<1405:POPWBI>2.0.CO;2), 1991.
- Hierro, R., Steiner, A. K., de la Torre, A., Alexander, P., Llamedo, P., and Cremades, P.: Orographic and convective gravity waves above the Alps and Andes Mountains during GPS radio occultation events - a case study, *Atmos. Meas. Tech.*, 11, 3523–3539, <https://doi.org/10.5194/amt-11-3523-2018>, 2018.
- 15 Hoffmann, L., Xue, X., and Alexander, M. J.: A global view of stratospheric gravity wave hotspots located with Atmospheric Infrared Sounder observations, *J. Geophys. Res.*, 118, 416–434, <https://doi.org/10.1029/2012JD018658>, 2013.
- Holton, J. R.: The role of gravity wave induced drag and diffusion in the momentum budget of the mesosphere, *J. Atmos. Sci.*, 39, 791–799, [https://doi.org/10.1175/1520-0469\(1982\)039<0791:TROGWI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1982)039<0791:TROGWI>2.0.CO;2), 1982.
- Jacobi, C., Fröhlich, K., and Pogoreltsev, A.: Quasi two-day-wave modulation of gravity wave flux and consequences for the planetary wave propagation in a simple circulation model, *J. Atmos. Sol.-Terr. Phys.*, 68, 283–292, <https://doi.org/10.1016/j.jastp.2005.01.017>, 2006.
- 20 Jacobi, C., Fröhlich, K., Portnyagin, Y., Merzlyakov, E., Solovjova, T., Makarov, N., Rees, D., Fahrutdinova, A., Guryanov, V., Fedorov, D., Korotyshkin, D., Forbes, J., Pogoreltsev, A., and Kürschner, D.: Semi-empirical model of middle atmosphere wind from the ground to the lower thermosphere, *Adv. Space Res.*, 43, 239–246, <https://doi.org/10.1016/j.asr.2008.05.011>, 2009.
- Jacobi, C., Lilienthal, F., Geißler, C., and Krug, A.: Long-term variability of mid-latitude mesosphere-lower thermosphere winds over Collm (51°N, 13°E), *J. Atmos. Sol.-Terr. Phys.*, 136, 174–186, <https://doi.org/10.1016/j.jastp.2015.05.006>, 2015.
- 25 Jakobs, H. J., Bischof, M., Ebel, A., and Speth, P.: Simulation of gravity wave effects under solstice conditions using a 3-D circulation model of the middle atmosphere, *J. Atmos. Terr. Phys.*, 48, 1203–1223, [https://doi.org/10.1016/0021-9169\(86\)90040-1](https://doi.org/10.1016/0021-9169(86)90040-1), 1986.
- Jiang, J. H., Wang, B., Goya, K., Hocke, K., Eckermann, S. D., Ma, J., Wu, D. L., and Read, W. J.: Geographical distribution and interseasonal variability of tropical deep convection: UARS MLS observations and analyses, *J. Geophys. Res.*, 109, <https://doi.org/10.1029/2003JD003756>, 2004.
- 30 Karami, K., Braesicke, P., Sinnhuber, M., and Versick, S.: On the climatological probability of the vertical propagation of stationary planetary waves, *Atmos. Chem. Phys.*, 16, 8447–8460, <https://doi.org/10.5194/acp-16-8447-2016>, 2016.
- Kirkwood, S., Mihalikova, M., Rao, T., and Satheesan, K.: Turbulence associated with mountain waves over Northern Scandinavia – a case study using the ESRAD VHF radar and the WRF mesoscale model, *Atmos. Chem. Phys.*, 10, 3583–3599, [https://doi.org/10.5194/acp-10-](https://doi.org/10.5194/acp-10-3583-2010)
- 35 3583-2010, 2010.
- Kumar, K., Ramkumar, T. K., and Krishnaiah, M.: Analysis of large-amplitude stratospheric mountain wave event observed from the AIRS and MLS sounders over the western Himalayan region, *J. Geophys. Res.*, 117, <https://doi.org/10.1029/2011JD017410>, 2012.



- Labitzke, K.: The Amplification of Height Wave 1 in January 1979: A Characteristic Precondition for the Major Warming in February, *Mon. Wea. Rev.*, 109, 983–989, [https://doi.org/10.1175/1520-0493\(1981\)109<0983:TAOHWI>2.0.CO;2](https://doi.org/10.1175/1520-0493(1981)109<0983:TAOHWI>2.0.CO;2), 1981.
- Li, Q., Graf, H. F., and Giorgetta, M. A.: Stationary planetary wave propagation in Northern hemisphere winter climatological analysis of the refractive index, *Atmos. Chem. Phys.*, 7, 183–200, <https://doi.org/10.5194/acp-7-183-2007>, 2007.
- 5 Lieberman, R. S., Riggan, D. M., and Siskind, D. E.: Stationary waves in the wintertime mesosphere: Evidence for gravity wave filtering by stratospheric planetary waves, *J. Geophys. Res.*, 118, 3139–3149, <https://doi.org/10.1002/jgrd.50319>, 2013.
- Lilienthal, F., Jacobi, C., Schmidt, T., de la Torre, A., and Alexander, P.: On the influence of zonal gravity wave distributions on the Southern Hemisphere winter circulation, *Ann. Geophys.*, 35, <https://doi.org/10.5194/angeo-35-785-2017>, 2017.
- Lilienthal, F., Jacobi, C., and Geißler, C.: Forcing mechanisms of the terdiurnal tide, *Atmos. Chem. Phys. Discuss.*, 18, 15 725–15 742, <https://doi.org/10.5194/acp-18-15725-2018>, 2018.
- 10 Lilly, D. K., Nicholls, J. M., Kennedy, P. J., Klemp, J. B., and Chervin, R. M.: Aircraft measurements of wave momentum flux over the Colorado Rocky mountains, *Quart. I. R. Meteor. Soc.*, 108, 625–641, <https://doi.org/10.1002/qj.49710845709>, 1982.
- Lindzen, R. S.: Turbulence and stress owing to gravity wave and tidal breakdown, *J. Geophys. Res.*, 86, 9707–9714, <https://doi.org/10.1029/JC086iC10p09707>, 1981.
- 15 Llamedo, P., de la Torre, A., Luna, P. A. D., Schmidt, T., and Wickert, J.: A gravity wave analysis near the Andes range from GPS radio occultation data and mesoscale numerical simulations: Two case studies, *Adv. Space Res.*, 44, 494–500, <https://doi.org/10.1016/j.asr.2009.04.023>, 2009.
- Matsuno, T.: A dynamical model of the stratospheric sudden warming, *J. Atmos. Sci.*, 28, 1479–1494, [https://doi.org/10.1175/1520-0469\(1971\)0282.0.CO;2](https://doi.org/10.1175/1520-0469(1971)0282.0.CO;2), 1971.
- 20 Matthias, V. and Ern, M.: On the origin of the mesospheric quasi-stationary planetary wave in the unusual Arctic winter 2015/2016, *Atmos. Chem. Phys.*, 18, 4803–4815, <https://doi.org/10.5194/acp-18-4803-2018>, 2018.
- Moffat-Griffin, T., Hibbins, R. E., Jarvis, M. J., and Colwell, S. R.: Seasonal variation of gravity wave activity in the lower stratosphere over an Antarctic Peninsula station, *J. Geophys. Res.*, 116, <https://doi.org/10.1029/2010JD015349>, 2010.
- Nastrom, G. D. and Fritts, D. C.: Sources of Mesoscale Variability of Gravity Waves. Part I: Topographic Excitation, *J. Atmos. Sci.*, 49, 101–110, [https://doi.org/10.1175/1520-0469\(1992\)049<0101:SOMVOG>2.0.CO;2](https://doi.org/10.1175/1520-0469(1992)049<0101:SOMVOG>2.0.CO;2), 1992.
- 25 Pišoft, P., Šácha und J. Mikšovský, P., Huszar, P., and Scherllin-Pirscher, B.: Revisiting internal gravity wave analysis using GPS RO density profiles: comparison with temperature profiles and application for wave field stability study, *Atmos. Meas. Tech.*, 11, 515–527, <https://doi.org/10.5194/amt-11-515-2018>, 2018.
- Plougonven, R. and Zhang, F.: Internal gravity waves from atmospheric jets and fronts, *Reviews of Geophysics*, American Geophysical Union, 52, 33–76, <https://doi.org/10.1002/2012RG000419>, 2014.
- 30 Plougonven, R., Hertzog, A., and Teitelbaum, H.: Observations and simulations of a large-amplitude mountain wave breaking over the Antarctic Peninsula, *J. Geophys. Res.*, 113, <https://doi.org/10.1029/2007JD009739>, 2008.
- Pogoreltsev, A. I., Vlasov, A. A., Fröhlich, K., and Jacobi, C.: Planetary waves in coupling the lower and upper atmosphere, *J. Atmos. Solar-Terr. Phys.*, 69, 2083–2101, <https://doi.org/10.1016/j.jastp.2007.05.014>, 2007.
- 35 Portnyagin, Y., Solovjova, T., Merzlyakov, E., Forbes, J., Palo, S., Ortland, D., Hocking, W., MacDougall, J., Thayaparan, T., Manson, A., Meek, C., Hoffmann, P., Singer, W., Mitchell, N., Pancheva, D., Igarashi, K., Murayama, Y., Jacobi, C., Kürschner, D., Fahrutdinova, A., Korotyshkin, D., Clark, R., Tailor, M., Franke, S., Fritts, D., Tsuda, T., Nakamura, T., Gurubaran, S., Rajaram, R., Vincent, R., Kovalam,

- S., Batista, P., Poole, G., Malinga, S., Fraser, G., Murphy, D., Riggan, D., Aso, T., and Tsutsumi, M.: Mesosphere/lower thermosphere prevailing wind model, *Adv. Space Res.*, 34, 1755–1762, <https://doi.org/10.1016/j.asr.2003.04.058>, 2004.
- Reid, I. M. and Vincent, R. A.: Measurements of mesospheric gravity wave momentum fluxes and mean flow accelerations at Adelaide, Australia, *J. Atmos. Terr. Phys.*, 49, 443–460, [https://doi.org/10.1016/0021-9169\(87\)90039-0](https://doi.org/10.1016/0021-9169(87)90039-0), 1987.
- 5 Šácha, P., Kuchar, A., Jacobi, C., and Pišoft, P.: Enhanced internal gravity wave activity and breaking over the Northeastern Pacific / Eastern Asian region, *Atmos. Chem. Phys.*, 15, 13 097–13 112, <https://doi.org/10.5194/acp-15-13097-2015>, 2015.
- Šácha, P., Lilienthal, F., Jacobi, C., and Pišoft, P.: Influence of the spatial distribution of gravity wave activity on the middle atmospheric circulation and transport, *Atmos. Chem. Phys.*, 16, 15 755–15 775, <https://doi.org/10.5194/acp-16-15755-2016>, 2016.
- Šácha, P., Miksovsky, J., and Pišoft, P.: Interannual variability in the gravity wave drag - vertical coupling and possible climate links, *Earth*  
 10 *Syst. Dynam.*, 9, 647–661, <https://doi.org/10.5194/esd-9-647-2018>, 2018.
- Schmidt, T., Alexander, P., and de la Torre, A.: Stratospheric gravity wave momentum flux from radio occultations, *J. Geophys. Res.*, 121, 4443–4467, <https://doi.org/10.1002/2015JD024135>, 2016.
- Smith, A. K.: The origin of stationary planetary waves in the upper mesosphere, *J. Atmos. Sci.*, 60, 3033–3041, [https://doi.org/10.1175/1520-0469\(2003\)060<3033:TOOSPW>2.0.CO;2](https://doi.org/10.1175/1520-0469(2003)060<3033:TOOSPW>2.0.CO;2), 2003.
- 15 Smith, R. B.: On Severe Downslope Winds, *J. Atmos. Sci.*, 42, 2597–2603, [https://doi.org/10.1175/1520-0469\(1985\)042<2597:OSDW>2.0.CO;2](https://doi.org/10.1175/1520-0469(1985)042<2597:OSDW>2.0.CO;2), 1985.
- Strobel, D. F.: Parameterization of the atmospheric heating rate from 15 to 120km due to O<sub>2</sub> and O<sub>3</sub> Absorption of solar radiation, *J. Geophys. Res.*, 83, 6225–6230, <https://doi.org/10.1029/JC083iC12p06225>, 1986.
- Swinbank, R. and Ortland, D. A.: Compilation of wind data for the Upper Atmosphere Research Satellite (UARS) Reference Atmosphere  
 20 Project, *J. Geophys. Res.*, 108, 4615, <https://doi.org/10.1029/2002JD003135>, 2003.
- Tsuda, T., Muayama, Y., Wiryosumarto, H., Harijono, S. W. B., and Kato, S.: Radiosonde observations of equatorial atmosphere dynamics over Indonesia. II: characteristics of gravity waves, *J. Geophys. Res.*, 99, 10.507–10.516, <https://doi.org/10.1029/94JD00354>, 1994.
- White, R. H., Battisti, D. S., and Sheshadri, A.: Orography and the Boreal Winter Stratosphere: The Importance of the Mongolian Mountains, *Geophys. Res. Lett.*, 45, 2088–2096, <https://doi.org/10.1002/2018GL077098>, 2018.
- 25 Wright, C. J. and Gille, J. C.: HIRDLS observations of gravity wave momentum fluxes over the monsoon regions, *J. Geophys. Res.*, 116, <https://doi.org/10.1029/2011JD015725>, 2011.
- Xiao, C., Hu, X., and Tian, J.: Global temperature stationary planetary waves extending from 20 to 120 km observed by TIMED/SABER, *J. Geophys. Res.*, 114, <https://doi.org/10.1029/2008JD011349>, 2009.

## Appendix A: Figures



**Figure 12.** Zonal mean EP flux (arrows) and divergence (isolines and shaded areas, dashed lines show negative values) of SPW 1. Shown is the difference between all H1-8 (a-h) simulations and the reference simulation (H1-8 – Ref) representing the last 30 days of the simulations. The EP flux is weighted by the density.