The role of the winter residual circulation in the summer mesopause regions in WACCM

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Abstract

High winter planetary wave activity warms the summer polar mesopause via a link between the two hemispheres. In a recent study carried out with the Kühlungsborn Mechanistic general Circulation Model (KMCM), it was shown that the net effect of this interhemispheric coupling mechanism is a cooling of the summer polar mesospheres and that this temperature response is tied to the strength of the gravity wave-driven winter mesospheric flow. We here reconfirm the hypothesis that the summer polar mesosphere would be substantially warmer without the circulation in the winter mesosphere, using the widely-used Whole Atmosphere Community Climate Model (WACCM). In addition, the role of the stratosphere in shaping the conditions of the summer polar mesosphere is investigated. Using composite analysis, we show that if winter gravity waves are absent, a weak stratospheric Brewer-Dobson circulation would lead to a warming of the summer mesosphere region instead of a cooling, and vice versa. This is opposing the temperature signal of the interhemispheric coupling in the mesosphere, in which a cold winter stratosphere goes together with a cold summer mesopause. We hereby strengthen the evidence that the equatorial mesospheric temperature response, driven by the winter gravity waves, is a crucial step in the interhemispheric coupling mechanism.

1 Introduction

The circulation in the mesosphere is driven by atmospheric gravity waves (GWs). These waves originate from the lower atmosphere and as they propagate upwards, they are filtered by the zonal wind in the stratosphere (e.g., Fritts and Alexander, 2003). Because of the decreasing density with altitude and as a result of energy conservation, the waves grow in amplitude. At certain altitudes, the waves – depending on their phase speeds relative to the background wind - become unstable and break. At the level of breaking, the waves deposit their momentum into the background flow, creating a drag on the zonal winds in the mesosphere, which establishes the pole-to-pole circulation (e.g. Lindzen, 1981; Holton, 1982,1983; Garcia and Solomon, 1985). This circulation drives the temperatures far away from the state of radiative balance, by adiabatically heating the winter mesopause and adiabatically cooling the summertime mesopause (Andrews et al., 1987; Haurwitz, 1961; Garcia and Solomon, 1985; Fritts and Alexander, 2003). The adiabatic cooling in the summer leads to temperatures sometimes lower than 130 K in the summer mesopause (Lübken et al.,1990). These low temperatures allow for the formation of thin ice clouds in the summer mesopause region, the so-called noctilucent clouds (NLCs).

Previous studies have shown that the summer polar mesosphere is influenced by the winter stratosphere via a chain of wave-mean flow interactions (e.g. Becker and Schmitz, 2003; Becker et al., 2004; Karlsson et al., 2009). This phenomenon, termed interhemispheric coupling (IHC), manifests itself as an anomaly of the zonal mean temperatures. Its pattern consists of a quadruple structure in the winter hemisphere with a warming (cooling) of the polar stratosphere and an associated cooling (warming) in the equatorial stratosphere. In the mesosphere, these anomalies are reversed: there is a cooling (warming) in the polar mesosphere, and an associated warming (cooling) in the equatorial region. The mesospheric warming (cooling) in the tropical region extends to the summer mesopause (see e.g. Körnich and Becker, 2010).

These anomalies are responses to different wave forcing in the winter hemisphere. To understand how these anomalies come about we have to understand the interhemispheric coupling mechanism. The mechanism, as discussed here, is for the case of a stronger winter residual circulation, but works the same for a weakening of this circulation (Karlsson et al., 2009). A stronger planetary wave (PW) forcing in the winter stratosphere yields a stronger stratospheric Brewer-Dobson circulation (BDC). This anomalously strong flow yields an anomalously cold stratospheric tropical region and a warm stratospheric winter pole, due to the downward control principle (Haynes et al. 1991).

Due to the eastward zonal flow in the winter stratosphere, GWs carrying westward momentum propagate relatively freely up through the mesosphere where they break. Therefore, in the winter mesosphere, the net drag from GWs momentum deposition is westward. When vertically propagating planetary waves break – also carrying westward momentum – in the stratosphere, the momentum deposited onto the mean flow decelerates the stratospheric westerly winter flow. To put it short, a weaker zonal stratospheric winter flow allows for the upward propagation of more GWs with an eastward phase speed, which, as they break reduces the westward wave drag (see Becker and Schmitz, 2003, for a more rigorous description).

This filtering effect of the zonal background flow on the GW propagation results in a reduction in strength of the winter-side mesospheric residual circulation when the BDC is stronger. The downward control principle now causes the mesospheric polar winter region to be anomalously cold and the tropical mesosphere to be anomalously warm (Becker and Schmitz, 2003, Becker et al., 2004; Körnich and Becker, 2009).

The critical step for IHC is the crossing of the temperature signal over the equator. The essential region is here the equatorial mesosphere. Central in the hypothesis of IHC is that the increase (or decrease) of the temperature in the tropical mesosphere modifies the temperature gradient between high and low latitudes in the summer mesosphere, which influences the zonal wind in the summer mesosphere, due to thermal wind balance (see e.g. Karlsson et al., 2009 and Karlsson and Becker, 2016).

The zonal wind change in the summer mesosphere modifies the breaking level of the summer-side GWs. In the case of a warming in the equatorial mesosphere – as when the BDC is strong -, the zonal wind is modified in such a way that the intrinsic wave speeds are reduced (e.g. Becker and Schmitz, 2003; Körnich and Becker, 2009). When the relative speed between the GWs and the zonal flow decreases, the GWs break at a lower altitude, thereby shifting down the GW drag per unit mass. The upper branch of the residual circulation also shifts downwards and along with this shift there is a reduction of adiabatic cooling, which causes a positive temperature anomaly in the summer mesosphere (Karlsson et al., 2009; Körnich and Becker, 2009; Karlsson and Becker, 2016). In the case of an equatorial mesospheric cooling, the response is the opposite: the relative difference between the zonal flow and the phase speeds of the gravity waves increase to that they break at a slightly higher altitude, with an anomalous cooling of the summer mesopause as a result.

The IHC pattern was first found using mechanistic models (Becker and Schmitz, 2003; Becker et al., 2004; Becker and Fritts, 2006), underpinned by observations of mesospheric conditions. The pattern was then found in observational data (e.g. Karlsson et al., 2007; Gumbel and Karlsson, 2011; Espy et al., 2011: de Wit et al., 2016), in the Whole Atmosphere Community Climate Model (WACCM: Sassi et al. 2004, Tan et al., 2012), in the Canadian Middle Atmosphere Model (CMAM: Karlsson et al. 2009), and in the high altitude analysis from the Navy Operational Global Atmospheric Prediction System- Advanced Level Physics High Altitude (NOGAPS-ALPHA) forecast/assimilating system (Siskind et al., 2011).

We saw that the temperature in the equatorial mesosphere is modified by the strength of the residual circulation in the winter mesosphere. Karlsson and Becker (2016) hypothesized that if the GW-driven winter residual circulation would not be present, the equatorial mesosphere would be warmer, which would lead to lower breaking levels of GWs in the summer hemisphere and a warmer summer mesosphere region. Analogically, an anomalously cold equatorial region would lead to an anomalously cold summer mesosphere region (e.g. Karlsson et al., 2009; Karlsson and Becker, 2016).

Becker and Karlsson (2016) showed that the equatorial mesosphere is substantially colder in July than it is in January, while the winter mesosphere is significantly warmer (see their Fig. 1). That means that the GWs break higher in the NH summer mesosphere than in the SH summer mesosphere, which is one possible reason for why the July summer polar mesosphere is colder than in the January summer polar mesosphere (e.g. Becker and Fritts, 2006; Karlsson et al., 2009). If – as hypothesized by Karlsson and Becker (2016) – the fundamental effect of the IHC is a cooling of the summer mesopauses, it would mean that the mechanism plays a more important role affecting the temperatures in the summer mesopause in the NH compared to that in the SH, since the weaker planetary wave activity in the SH results in an increased gravity wave drag and a strengthening of mesospheric poleward flow in the winter mesosphere. The equatorial mesosphere is adiabatically cooled more efficiently than when the winter mesospheric circulation is weak.

Karlsson and Becker (2016) hypothesized that in the absence of the equator-to-pole flow in the SH winter, the summer mesopause in the NH would be considerably warmer. Moreover, removing the mesospheric residual circulation in the NH winter would not have as high impact on the SH summer mesopause. To test the hypothesis, they used the KMCM to compare control simulations to runs without GWs in the winter mesosphere. The predicted responses were confirmed, and the results were also backed up by correlation studies using the Canadian Middle Atmosphere Model (CMAM30).

Since IHC is controversial, we find it important to use as many tools as possible to test – and to underpin - our arguments. In this study, the well-established WACCM, described in section 2.1 below, is used to endorse the results obtained with the not as widely-used – yet high-performing – KMCM. WACCM is in some aspects a more comprehensive model than KMCM. For example, a major difference is that WACCM contains interactive chemistry in the middle atmosphere, while KMCM does not. WACCM also uses a different parameterization for non-orographic GWs than KMCM. KMCM uses a simplified dynamical core and convection scheme as compared to WACCM. For details about the KMCM see e.g. Becker et al., 2015. The WACCM is described in section 2.

In section 3, we discuss the effect of removing the gravity waves in the winter hemisphere on the summer mesosphere region in WACCM. We also investigate the consequences for noctilucent clouds, formed in the mesopause region. Therefore, we implement a basic cloud parameterization, as described in Section 2.2. The Whole Atmosphere Community Climate Model (WACCM) results from comparing runs with and without winter GWs are presented in Section 3.

As an important complement to the study carried out by Karlsson and Becker (2016), we here examine the role of the summer stratosphere in shaping the conditions of the NH summer polar mesosphere when the winter mesospheric flow is absent. We focus on the effect that the zonal wind in the summer stratosphere has, and study if and how the PW activity in the winter affects the summer polar mesosphere. These results are presented in Section 3.1.

Our conclusions are summarized in Section 4. Since the IHC mechanism has a more robust signal in the SH winter – NH summer, we choose to focus particularly on this period, namely July. Nevertheless, results from January are also shown for comparisons and for further discussion.

2 Method

2.1 Model

The Whole Atmosphere Community Climate Model (WACCM) is a so-called “high-top” chemistry-climate model, which spans the range of altitude from the Earth’s surface to an altitude of about 140 km. WACCM has 66 vertical levels of a resolution of ~1.1 km in the troposphere above the boundary layer, 1.1-1.4 km in the lower stratosphere, 1.75 km at the stratosphere and 3.5 km above 65 km. The horizontal resolution is 1.9° latitude by 2.5° longitude (Marsh et al, 2013).

The model is a component of the Community Earth System Model (CESM), which is a group of model components at the National Center for Atmospheric

Research (NCAR). WACCM is a superset of the Community Atmospheric Model version 4 (CAM4) and as such it includes all the physical parameterizations of CAM4 (Neale et al., 2013).

WACCM includes parameterized non-orographic gravity waves, which are generated by frontal systems and convection (Richter et al., 2010). The orographic GW parameterization is based on McFarlane (1987), while the nonorographic GW propagation parameterization is based the formulation by Lindzen (1981).

In this study, The F_2000_WACCM (FW) compset of the model is used, i.e. the model assumes present day conditions. There is no forcing applied: the model runs a perpetual year 2000. Our results are based on a control run and perturbation runs. In the control run, the winter side residual circulation is included. In the perturbation runs, the equator-to-pole flow is removed by turning off both the orographic and the non-orographic gravity waves. It should however be noted that even though the GWs are turned off, there are still some resolved waves, such as inertial gravity waves and planetary waves that drive a weak meridional circulation. The model is run for 30 years.

3 Results and discussion

To investigate the effect of the winter residual circulation on the summer mesopause, we compare the control run, which includes the winter equator-to-pole circulation, with the perturbation runs. In the perturbation runs, the equator-to-pole flow is removed by turning off the parameterized gravity waves in the winter hemisphere. The resolved waves, such as tides, inertial gravity waves and planetary waves are still there and drive a weak poleward flow, as already described in section 2.1.

We start by investigating the case for the NH summer (July) with the GWs turned off for the SH, where it is winter. Figure 1 shows the difference in zonal-mean temperature, zonal wind and gravity wave drag for July as a function of latitude and altitude, between the control run and the perturbation run: the run without the GWs in the winter minus the run with the GWs in the SH.

Figure 1.

From Fig. 1, it is clear that there is a considerable increase in temperature in the NH summer mesopause region in the case for which there is no equator-to-pole flow in the SH winter. This change in temperature in the summer polar mesosphere can be understood as a result of changes in the wave-mean flow interactions. Without the GWs in the SH winter, the winter stratosphere and lower mesosphere are colder. This is because GWs in the winter hemisphere drive downwelling, adiabatically heating these regions (e.g. Karlsson et al., 2009).

From Fig. 1 it can also be seen that there is a significant warming in the equatorial stratosphere in the case where there are no GWs in the winter hemisphere, indicating a weakening of the BDC. We suggest that this could be due to a redistribution of PW momentum drag in the winter stratosphere: as the zonal flow is no longer reversed in the mesosphere by GW-drag, the breaking levels of the westward propagating planetary waves are shifted upwards. Hence, the PW drag could be distributed over a wider altitude range. Another contributor to a decrease in the BDC is the removal of the orographic GWs, which act as PWs on the zonal flow in the winter stratosphere (see e.g. Karlsson and Becker, 2016; their figure 7).

Turning off the gravity waves in winter hemisphere, changes the meridional temperature gradient in the winter hemisphere, as the equatorial mesosphere will be warmer. This tropical temperature response changes the meridional temperature gradient in the summer mesosphere, and thereby – via thermal wind balance - the zonal mesospheric winds: the westward jet will be weaker. It is also clear that the zonal flow at high latitudes accelerates for the case for which there is no equator-to-pole flow in the SH winter. These findings correspond with what is found in Karlsson and Becker (2016).

The weaker jet leads in turn to lower GW levels and weaker GW drag over 45°N-70°N above a pressure level of 0.02 hPa as can be seen in Fig. 1. This causes the summer polar mesopause to be considerably warmer. The temperature increase in the summer polar mesopause region, which is now loosely defined to be between 61°N - 90°N and 0.01 - 0.002 hPa, is approximately 16 degrees. In a radiation-driven atmosphere the temperature in the NH NLC region is about 210-220 K, much higher than the temperature both with and without the GWs in the SH.

When we compare our results with the results in Karlsson and Becker (2016, their figure 3), we observe there are some quantitative discrepancies in the structure of the responses. For example, Karlsson and Becker (2016) found that removing the winter GWs resulted in a warming of the mesosphere globally, although the response was strongest in the polar mesopause region. They attributed that the warming over the equatorial and winter mesosphere to the effect that GWs have on tides: when GWs are absent, the tidal response is enhanced. The same behavior is not found in WACCM - in fact, the equatorial upper mesosphere is anomalously cooler when the GWs are removed. These differences could perhaps be explained by for example the different gravity wave parameterization of non-orographic GWs, the different dynamical cores between the models and the presence of interactive chemistry in the middle atmosphere in WACCM. However, the qualitative response of the temperature and zonal wind change due to turning of the GWs in the SH corresponds well with the results from the KMCM as well as with our hypothesis.

It can also be seen that like in the KMCM model, the zonal wind and temperature in summer stratosphere region change only slightly in the perturbation runs as compared to the control runs. We deem that anomalous GW filtering effects from the lower down in the summer stratosphere, which could affect the results, are unlikely to contribute substantially to the temperature change in the summer mesosphere. We come back to this question in the next paragraph 3.1.

To investigate the IHC mechanism further, we also show the correlation and covariance, which also provides information about the amplitude of the variability, between the temperature in the winter stratosphere in July (1-10 hPa, 60°S-40°S) and the temperatures in the rest of the atmosphere in the same month. The latitude and altitude ranges chosen for July is the region where the SH winter stratosphere variability is best captured (see Karlsson and Becker, 2016; their figure 9). This is related to the relatively weak PW forcing in the SH – the BDC is not reaching all the way to the polar region (Kuroda and Kodera, 2001).

We show the correlation and covariance fields for both the cases with and without GWs in the SH winter hemisphere.

Figure 2

In the correlation and covariance fields of the control run, the temperature in the winter stratosphere is positively correlated with the temperature in the equatorial mesosphere and the summer mesopause region. Comparing the results show in Figure 2 (upper left) to Figure 8e in Karlsson and Becker (2016), it can be seen that the correlation coefficients are of similar magnitudes, but the spatial responses differ in altitude and in latitudinal extent: whereas the correlation signal is significant in the CMAM30 July high latitude summer mesopause, the WACCM July response reaches only the lowermost latitudes (about 50°N in latitude).

If the GWs are removed in the winter hemisphere, the temperature in the summer mesopause region anti-correlates with the temperature in the winter stratosphere. Also, the temperature in the equatorial mesosphere does no longer correlate and co-vary significantly with the temperature in the winter hemisphere, in agreement with the results of Karlsson and Becker (2016).

Until now, we investigated the influence of the SH winter residual circulation on the NH summer mesopause in July. Now, we will also investigate the effect that the NH winter residual circulation has on the SH summer mesosphere in January. We discussed earlier that this effect will be smaller as compared to the effect of the SH winter residual circulation on the NH summer mesosphere in July. Figure 3 shows the difference in zonal-mean temperature, zonal wind and gravity wave drag for January as a function of latitude and altitude, between the control run and the perturbation run: the run without the GWs in the NH winter hemisphere minus the run with the GWs in the NH winter hemisphere (similar to Fig. 1).

Figure 3.

From Fig. 3, it can be observed that there is no statistically significant temperature change in the SH summer polar mesopause region in the case for which there is no equator-to-pole flow in the NH winter. Without the GWs in the winter hemisphere, the winter stratosphere and lower mesosphere are colder, as in the July case. There is a change in zonal wind at high southern latitudes, but there is no clear statistical significant increase. These findings correspond with what is hypothesized in the introduction: taking away the GWs in the NH winter will have a less effect on the SH summer mesopause than taking away the GWs in the SH winter on the NH summer mesopause.

This is due to the variable nature of the winter stratosphere zonal flow in the NH, which oscillates between being weak and strong. As a result, the January equatorial mesosphere is modified continuously: it varies between being cooled and warmed by the winter mesospheric residual flow. In July, on the other hand, the equatorial region is continuously cooled by the strong mesospheric residual flow in the SH winter. Hence, the interhemispheric coupling mechanism gives a plausible explanation to why the July summer mesosphere region is considerably colder than the one in January.

Comparison between the responses found using WACCM with those found with KMCM (Karlsson and Becker, 2016, their Fig. 3), shows that the temperature change is larger and extends all the way to the summer pole in KMCM, while this is not the case in WACCM. Moreover, the change in temperature in this region is not statically significant in WACCM. The differences in temperature and zonal wind responses are larger in January than in July when comparing the results of WACCM with that of KMCM. Nevertheless, the qualitative structure of the temperature and zonal wind change due to turning of the winter GWs corresponds convincingly well.

In Fig. 4, we show the correlation and covariance between the temperature in the winter stratosphere in January (1-10 hPa, 60°N-80°N) and the temperatures in the rest of the atmosphere in the same month for both the cases with and without GWs in the NH winter hemisphere.

Figure 4

The general pattern in January for the correlation and covariance for both the control run and the run without GWs in the winter hemisphere is very similar to the pattern in July. However, the correlation and covariance in the summer mesosphere with the temperatures in the winter stratosphere are not statistically significant. This can be understood, as the variability in the SH summer mesopause region in January is much higher. It is seen that in both hemispheres, the temperature in the equatorial mesosphere correlates statistically significant with the temperatures in the winter stratosphere for the control case, but not for the case without the GWs in the winter hemisphere.

IHC has hitherto primarily been seen as a mode of internal variability giving rise to a warming of the summer polar mesopause region. These results presented here and in Karlsson and Becker (2016) show the more fundamental role of interhemispheric coupling; the mechanism has a net cooling effect on the summer polar mesosphere. This study reconfirms this fundamental role of the IHC mechanism and strengthens the evidence that the equatorial mesospheric temperature response is the crucial step in the interhemispheric coupling mechanism.

3.1 The role of the summer stratosphere region

The BDC is modifying in the summer stratospheric meridional temperature gradient. Hence, filtering effects taking place below the mesosphere may seem like an additional - or alternative – mechanism to the response observed in the summer mesopause. In this section, we will discuss why this cannot be the case. We focus again mostly on the NH summer polar mesosphere region.

In Fig. 1, it is seen that if there are GWs in the SH winter hemisphere the temperature in the winter stratosphere is positively correlated with the temperature in the NH summer polar mesosphere. This means that for a stronger Brewer-Dobson circulation (BDC) and the resulting anomalously warm (cold) temperatures in the stratosphere at 40°- 60°S, there will be also an anomalously warm (cold) temperature in the summer polar mesosphere.

A strong or weak BDC results in a temperature change in the equatorial mesosphere, which changes the meridional temperature gradient in the summer mesosphere. As a result of the change in strength of the BDC, there is a change in the meridional temperature gradient as well, however, this gradient will have an opposite sign, as can be seen from Fig 1. As pointed out by Karlsson et al. (2009), the expected GW filtering effect of this stratospheric temperature gradient would oppose that of the mesospheric temperature gradient.

This can been shown clearly with the mesospheric winter residual circulation being out of play. From Fig. 2, it can be seen that anomalously low temperatures in the SH winter stratosphere, indicating a weak Brewer-Dobson circulation, without the GWs in the winter lead to a warming in the NH summer mesopause region, instead of a cooling as observed in the case where there are GWs in the SH winter hemisphere.

We hypothesize that this opposing signal is – in the absence of a mesospheric residual flow in the winter - caused by a modulation of the meridional temperature gradient in the summer stratosphere, inferred by the BDC.

To strengthen our arguments, we plot the vertical profiles of the zonal wind, GW drag between 45°N-55°N and the temperatures between 70°N-90°N in July. These profiles are shown for both high and low temperatures in the winter stratosphere (1-10 hPa, 60°S-40°S). The differences between the cases with anomalously low and high temperatures are also plotted.

Figure 5

From Fig. 5, it is clear for a weak Brewer-Dobson circulation, and therefore anomalously low temperatures in the SH winter stratosphere, the zonal winds in the stratosphere are less strongly westwards. This leads to a weaker GW drag and a warmer NH summer mesopause region.

We hereby suggest that without GWs in the SH winter hemisphere, it would be the variability in the NH summer stratosphere caused by the winter-side BDC that would have the major influence on the temperatures in the NH summer mesopause. A weaker (stronger) Brewer-Dobson circulation would lead to a change in the temperature gradient in the summer stratopause, which would lead to a cooling (warming) instead of the warming (cooling) associated with interhemispheric coupling.

The same is true for the effect of the SH summer stratosphere on the SH summer mesosphere in January. The profiles for the southern hemisphere in January are very similar to the profiles for the northern hemisphere in July, see figure 6.

Figure 6.

This means that in both the northern and summer hemisphere, a weaker (stronger) Brewer-Dobson circulation leads to a change in the temperature gradient in the summer stratopause, which leads to a warming (cooling) instead of the cooling (warming) that is associated with interhemispheric coupling.

4 Conclusions

In this study, the interhemispheric coupling mechanism and the role of the summer stratosphere in shaping the conditions of the summer polar mesosphere have been investigated. We have used the widely used WACCM model to reconfirm the hypothesis of Karlsson and Becker (2016) that the summer polar mesosphere would be substantially warmer without the gravity wave-driven residual circulation in the winter. We find, in accordance with the previous study, that the interhemispheric coupling mechanism has a net cooling effect on the summer polar mesospheres. We also find that the mechanism plays a more important role affecting the temperatures in the summer mesopause in the NH compared to that in the SH.

We have also investigated the role of the summer stratosphere in shaping the conditions of the summer polar mesosphere. It is shown that without the winter mesospheric residual circulation, the variability in the summer polar mesosphere is determined by the temperature gradient in the summer stratosphere below, which is modulated by the strength of the BDC. We have found that for both the northern and the southern hemisphere, in the absence of winter gravity waves, a weak Brewer-Dobson circulation would lead to a warming of the summer mesosphere region. The temperature signal of the interhemispheric coupling mechanism is opposite: in this case a weak Brewer-Dobson circulation, the summer mesosphere region is cooled. This confirms the idea that it is the equatorial mesosphere that is governing the temperatures in the summer mesopause regions, rather than processes in the summer stratosphere.

References

Andrews, D.G., Holton, J.R., Leovy, C.B.: Middle atmosphere dynamics, Academic Press, United States of America, 1987.

Becker E. and Fritts, D.C.: Enhanced gravity-wave activity and interhemispheric coupling during the MaCWAVE/MIDAS northern summer program 2002, Ann. Geophys., 24, 1175-1188, doi:10.5194/angeo-24-1175-2006, 2006.

Becker, E. and Schmitz, G.: Climatological effects of orography and land-sea contrasts on the gravity wave-driven circulation of the mesosphere, J. Atmos. Sci., 60, 103-118, doi:.1175/1520-0469(2003)060<0103:CEOOAL>2.0.CO;2, 2003.

Becker, E., R. Knöpfel and F.-J. Lübken, 2015: Dynamically induced hemispheric differences in the seasonal cycle of the summer polar mesopause. J. Atmos. Solar-Terr. Phys., 129, 128-141, doi:10.1016/j.jastp.2015.04.014.

Becker, E., Müllermann, A., Lübken, F.-J., Körnich, H., Hoffmann, P., Rapp, M.: High Rossby-wave activity in austral winter 2002: Modulation of the general circulation of the MLT during the MaCWAVE/MIDAS northern summer program, Geophys. Res. Lett., 31, L24S03, doi:10.1029/2004GL019615, 2004.

Christensen, O.M., Benze, S., Eriksson, P., Gumbel, J., Megner, L., Murtagh, D.P.: The relationship between Polar Mesospheric Clouds and their background atmosphere as observed by Odin-SMR and Odin-OSIRIS, Atmos. Chem. Phys., 16, 12587-12600, doi:10.5194/acp-16-12587-2016, 2016.

De Wit, R.J., Janches, D., Fritts, D.C. , Hibbins, R.E.: QBO modulation of the mesopause gravity wave momentum flux over Tierra del Fuego, Geophys. Res. Lett., 43, 4094-4055, doi:10.1002/2016GL068599, 2016.

Espy, P.J., Ochoa Fernández, S., Forkman, P., Murtagh, D., Stegman, J.:

The role of the QBO in the inter-hemispheric coupling of summer mesospheric temperatures, Atmos. Chem. Phys., 11, doi:10.5194/acp-11-495-2011, 495-502, 2011.

Fritts, D. C. and Alexander, M.J.: Gravity wave dynamics and effects in the middle atmosphere, Rev. Geophys., 41, 1003, doi:10.1029/2001RG000106, 2003.

Garcia, R.R., Solomon, S.: The effect of breaking gravity waves on the dynamics and chemical composition of the mesosphere and lower thermosphere, J. Geophys. Res., 90, D2, 3850-3868, 10.1029/JD090iD02p03850, 1985.

Gumbel, J., Karlsson, B.: Intra- and inter-hemispheric coupling effects on the polar summer mesosphere, Geophys. Res. Lett., 38, L14804, doi:10.1029/2011GL047968, 2011.

Haurwitz, B.: Frictional effects and the meridional circulation in the mesosphere, J. Geophys. Res., 66, 8, doi:10.1029/JZ066i008p02381, 1961.

Haynes, P.H., Marks, C.J., McIntyre, M.E., Shepherd, T.G., Shine, K.P.: On the “downward control” of extratropical diabatic circulations by eddy-induced mean zonal forces, J. Atmos. Sci., 48, 651-678, 10.1175/1520-0469(1991)048<0651:OTCOED>2.0.CO;2, 1991.

Hervig, M.E., Stevens, M.H., Gordley, L.L., Deaver, L.E., Russell, J.M., Bailey, S.M.: Relationships between polar mesospheric clouds, temperature, and water vapor from Solar Occultation for Ice Experiment (SOFIE) observations, J. Geophys. Res., 114, D20203, doi:10.1029/2009JD012302, 2009.

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, doi:10.1175/1520-0469(1982)039<0791:TROGWI>2.0.CO;2, 1982.

Holton, J.R.: The influence of gravity wave breaking on the general circulation of the middle atmosphere, J. Atmos. Sci., 40, 2497-2507, doi:10.1175/1520-0469(1983)040<2497:TIOGWB>2.0.CO;2, 1983.

Karlsson, B., Körnich, H., Gumbel, J.: Evidence for interhemispheric stratosphere-mesosphere coupling derived from noctilucent cloud properties, Geophys. Res. Lett., 34, L16806, doi:10.1029/2007GL030282, 2007.

Karlsson, B., McLandress, C., Shepherd, T.G.: Inter-hemispheric mesospheric coupling in a comprehensive middle atmosphere model, J. Atmos. Sol.-Terr. Phys., 71, 3-4, 518-530, doi:10.1016/j.jastp.2008.08.006, 2009.

Karlsson, B., Becker, E.: How does interhemispheric coupling contribute to cool down the summer polar mesosphere?, 29, 8807-8821, doi:10.1175/JCLI-D-16-0231.1, J. Climate, 2016.

Körnich, H. and Becker, E.: A simple model for the interhemispheric coupling of the middle atmosphere circulation, Adv. in Space Res., 45, 5, 661-668, doi:10.1016/j.asr.2009.11.001, 2010.

Kuroda, Y., and K. Kodera: Variability of the polar night jet in the Northern and Southern Hemispheres, J. Geophys. Res., 106(D18),20,703–20, 713, 2001.

Lindzen, R.S.: Turbulence stress owing to gravity wave and tidal breakdown, J. Geophys. Res., 86, C10, 9707-9714, 10.1029/JC086iC10p09707, 1981.

Lübken, F.-J., Von Zahn, U., Manson, A., Meek, C., Hoppe, U.-P., Schmidlin, F.J., Stegman, J., Murtagh, D.P., Rüster, R., Schmitz, G., Widdel, H.-U., Espy, P.: Mean state densities, temperatures and winds during the MAC/SINE and MAC/EPSILON campaigns, J. Atmos. Sol.-Terr. Phys., 52, 10-11, 955-970, https://doi.org/10.1016/0021-9169(90)90027-K, 1990.

Richter, J. H., Sassi, F., Garcia, R.R.: Toward a physically based gravity wave source parameterization in a general circulation model, J. Atmos. Sci., 67, 136–156, DOI: 10.1175/2009JAS3112.1, 2010.

Rong, P., Russell, J. M., Gordley, L. L., Hervig, M. E., Deaver, L., Bernath, P. F., Walker, K. A.: Validation of v1.022 mesospheric water vapor observed by the Solar Occultation for Ice Experiment instrument on the Aeronomy of Ice in the Mesosphere satellite, J. Geophys. Res., 115, D24314, doi:10.1029/2010JD014269, 2010.

Marsh, D. R., Mills, M.J., Kinnison, D.E., Lamarque, J.F., Calvo, N., Polvani, L.M.: Climate change from 1850 to 2005 simulated in CESM1(WACCM), 73727391, J. Climate, 26, 19, doi:10.1175/JCLI-D-12-00558.1, 2013.

McFarlane, N. A.: The effect of orographically excited wave drag on the general circulation of the lower stratosphere and troposphere, J. Atmos. Sci., 44, 1775–1800, 10.1175/1520-0469(1987)044<1775:TEOOEG>2.0.CO;2, 1987.

Megner, L.: Minimal impact of condensation nuclei characteristics on observable mesospheric ice properties, J. Atmos. Sol.-Terr. Phys, 73, 14-15, 2184-2191, doi:10.1016/j.jastp.2010.08.006, 2011.

Merkel, A.W., Marsh, D.R., Gettelman, A., Jensen, E.J: On the relationship of polar mesospheric cloud ice water content particle radius and mesospheric temperature and its use in multi-dimensional model, Atmos. Chem. Phys., 9, 8889-8901, doi:10.5194/acp-9-8889-2009, 2009.

Murphy, D. M. and Koop, T.: Review of the vapor pressure of ice and super-cooled water for atmospheric applications, Q. J. R. Meteorol. Soc., 131, 1539-1565, doi:10.1256/qj.04.94, 2005.

Neale, R., Richter, J., Park, S., Lauritzen, P., Vavrus, S., Rasch, P., Zhang, M.: The mean climate of the Community Atmosphere Model (CAM4) in forced

SST and fully coupled experiments, J. Climate, 26, 5150–5168, doi:10.1175/JCLI-D-12-00236.1, 2013.

Sassi, F., Kinnison, D., Boville, B.A., Garcia, R.R., Roble, R.: Effect of El Niño- Southern Oscillation on the dynamical, thermal, chemical structure of the middle atmosphere, J. Geophys. Res., 109, D17108, doi:10.1029/2003JD004434, 2004.

Siskind, D. E., Stevens, M.H., Hervig, M., Sassi, F., Hoppel, K., Englert, C.R., Kochenas, A.J., Consequences of recent Southern Hemisphere winter variability on polar mesospheric clouds, J. Atmos. Sol. Terr. Phys., 73, 2013–2021, 10.1016/j.jastp.2011.06.014, 2011.

Tan, B., Chu, X., Liu, H.-L.: Yamashita, C., Russell III, J.M.: Zonal-mean global teleconnections from 15 to 110 km derived from SABER and WACCM, J. Geophys. Res., 117, D10106, doi:10.1029/2011JD016750, 2012.

Thomas, G. E., Olivero, J.J.: Climatology of polar mesospheric clouds 2. Further analysis of Solar Mesosphere Explorer data, J. Geophys. Res., 96, 14673-14681, doi:10.1029/JD094iD12p14673, 1989.

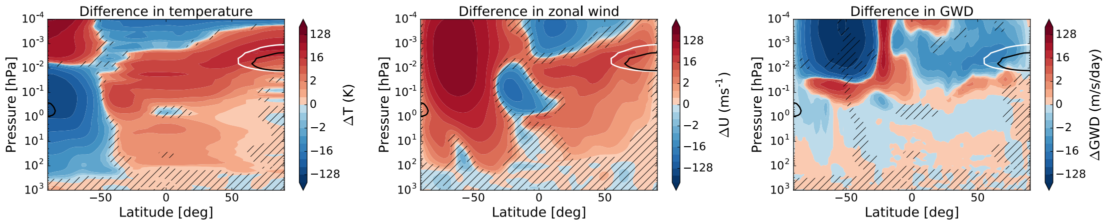

Fig.1. The difference in zonal-mean temperature (left) and zonal-mean zonal wind (right) for July: [run without winter GWs] minus [control run]. The white contour indicates the summer polar mesopause region where the temperatures are below 150 K for the control run. The black contour indicates the region where the temperature is below 150 K for the run without the GWs in winter. The shaded areas are regions where the data doesn't reach a confidence level of 95%.

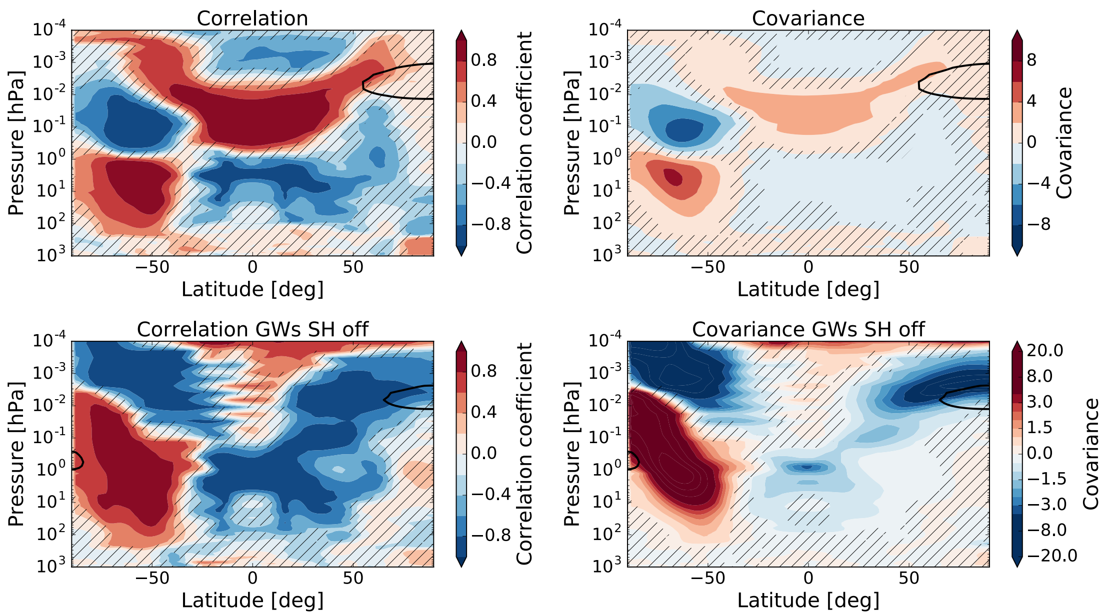

Fig. 2. The correlation (left) and covariance (right) between the temperature in the winter stratosphere (1-10 hPa, 60°S-40°S) and the temperatures in the rest of the atmosphere in July for the control run (first row) and run without GWs in the winter hemisphere (bottom row). The dotted areas are regions where the correlation has a p-value < 0.05. The black 150 K-contour indicates the polar mesopause region.

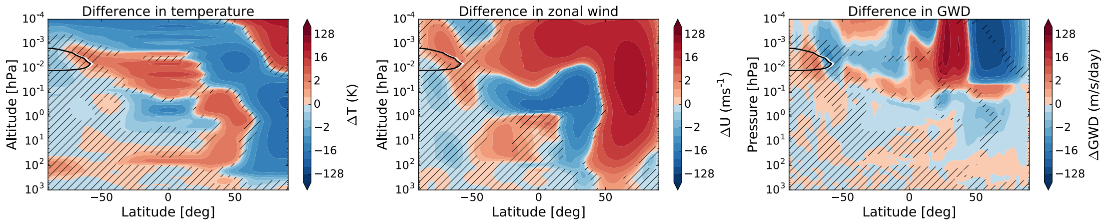

Fig. 3. Same as Figure 1, but for January.

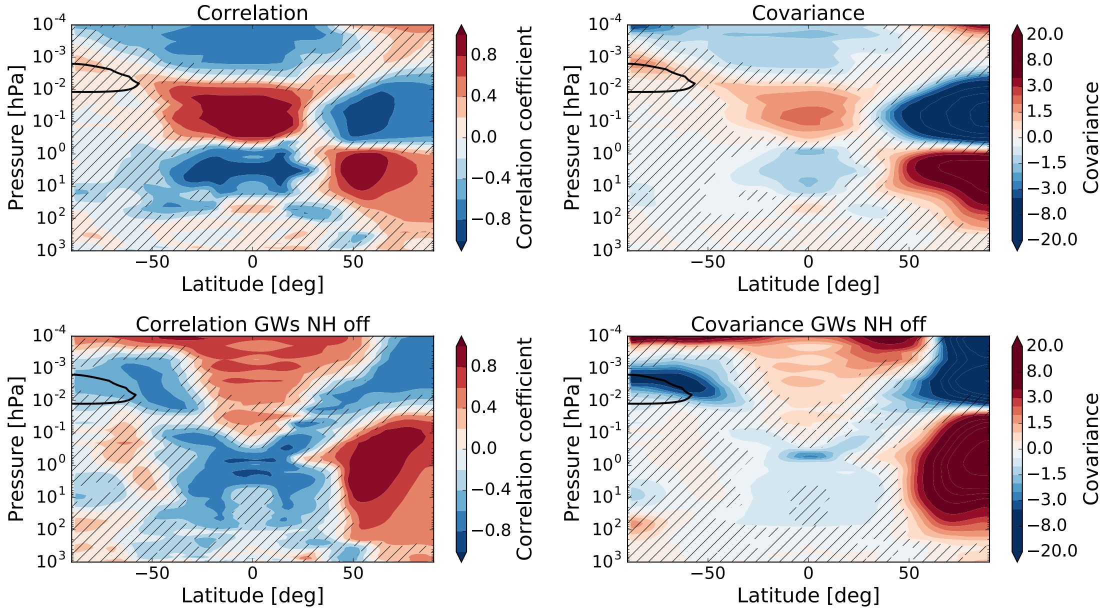

Fig. 4. The correlation (left) and covariance (right) between the temperature in the winter stratosphere (1-10 hPa, 40°N-60°N) and the temperatures in the rest of the atmosphere in January for the control run (first row) and run without GWs in the winter hemisphere (bottom row). The black 150 K-contour indicates the polar mesopause region. The dotted areas are regions where the correlation has a p-value < 0.05.

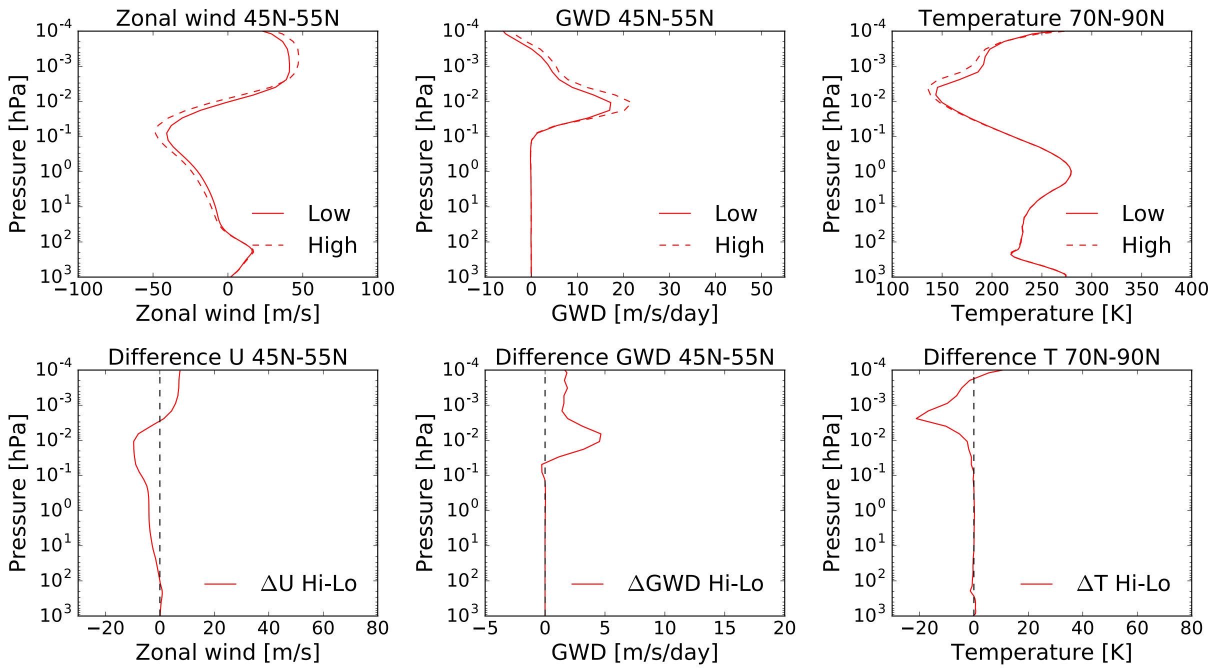

Fig. 5. The July zonal wind (left) and the GW drag (middle) between 45°-55°N and the temperature (right) between 70-90°N for anomalously low and high temperatures in the winter stratosphere (1-10 hPa, 60°S - 40°S) (first row) and the differences between them (second row), for the case where there are no GWs in the winter hemisphere. The red continuous lines show the results for anomalously low temperatures, the red dotted lines show the results for the anomalously high temperatures.

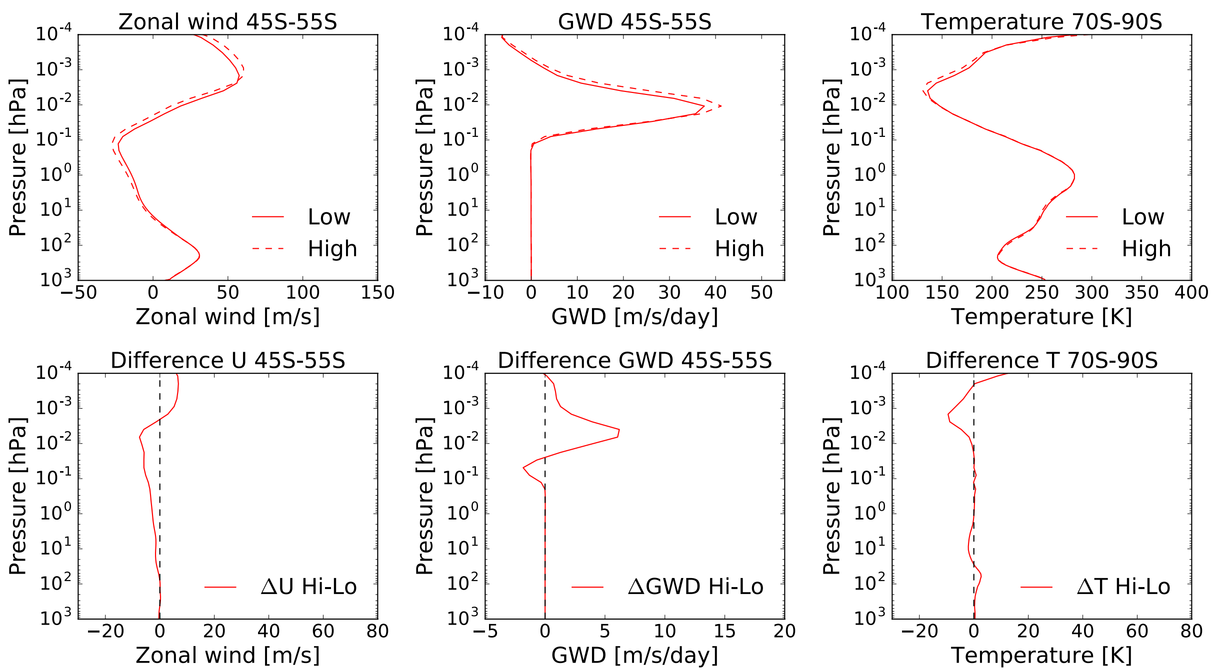

Fig. 6. The January zonal wind (left) and the GW drag (middle) between 45°-55°S and the temperature (right) between 70°S-90°S for anomalously low and high temperatures in the winter stratosphere (1-10 hPa, 40°N - 60°N) (first row) and the differences between them (second row), for the case where there are no GWs in the winter hemisphere. The red continuous lines show the results for anomalously low temperatures, the red dotted lines show the results for the anomalously high temperatures.