How Does Interhemispheric Coupling Contribute to Cool Down the Summer Polar Mesosphere?

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ABSTRACT

Interhemispheric coupling is commonly associated with events of high planetary wave activity in the winter stratosphere triggering a heating of the polar mesopause region in the opposite hemisphere. Here, a more fundamental role that this mechanism plays in the absence of planetary wave variability is highlighted. This study focuses directly on the mesospheric part of the coupling chain, which is induced by the gravity wave drag in the winter mesosphere. To investigate the effect that the winter residual flow has on the summertime high-latitude upwelling, the K"uhlungsborn Mechanistic General Circulation Model (KMCM) is used to compare a control simulation to runs where the parameterized gravity waves are removed from the winter hemisphere. The model response in the summer mesosphere reveals that the winter mesospheric residual circulation fosters a net (and substantial) cooling of the summer polar mesopause. These results offer an extension of the current view of interhemispheric coupling: from a mode of internal variability to a constant, gravity wave–driven phenomenon that is modulated by planetary wave activity.

1. Introduction

Zonal-mean hemispheric differences in the troposphere and (winter) stratosphere are mainly attributed to stronger planetary Rossby waves in the Northern Hemisphere (NH), causing the stratospheric polar night jet to be, on average, weaker, warmer, and more variable than in the Southern Hemisphere (SH). The summer stratospheres, on the other hand, are calm in both hemispheres since easterlies suppress the vertical propagation of planetary waves. At higher altitudes, however, the picture is different: it is the southern mesosphere during spring and summer that is warmer and more variable than its northern counterpart (e.g., Hervig and Siskind 2006; Bailey et al. 2007; Gumbel and Karlsson 2011). Unlike for the hemispheric differences in the troposphere and stratosphere, there is no consensus at this point about the mechanisms behind hemispheric differences between the polar summer mesopauses: it is still an open question. While the variability of the southern summer mesosphere has been a subject of investigations (e.g., Siskind and McCormack 2014; Becker et al. 2015), the relative calmness and coldness of the Northern Hemispheric summer mesosphere has not been considered much on its own. In this study, we want to draw the attention to the July summer polar mesospheric mean state.

Becker and Schmitz (2003) found that changes in planetary wave forcing in the winter stratosphere lead to a modulation of the gravity wave–driven branch of the residual circulation, not only in the winter mesosphere, but also in the summer-side mesosphere. This particular coupling—coined interhemispheric coupling by Becker et al. (2004)—can be understood as a chain of wave–mean flow interactions that is activated in the winter stratosphere by quasi-stationary planetary waves (PWs), which in turn are forced in the winter troposphere. The PWs decelerate the mean zonal wind in the winter stratosphere, and, as a consequence, they modify...
the levels at which nonorographic gravity waves (GWs) become unstable and break (see, e.g., Becker and Schmitz 2003). This modification results in a reduction of the GW drag in the winter mesosphere, which in turn affects the residual flow in the mesosphere globally.

The interhemispheric coupling mechanism (IHC) may be interpreted in the framework of the steady transformed Eulerian mean equations (e.g., Andrews et al. 1987, chapter 3). The signal is triggered in the winter mesosphere by the gravity wave–driven poleward residual circulation, which modifies the mesospheric temperature both at high latitudes (downwelling and adiabatic heating) and at the equator (upwelling and adiabatic cooling). The equatorial temperature response is most crucial for IHC because it modifies the temperature gradient between high and low latitudes in the summer mesosphere. The associated change in wind shear, via thermal wind balance, then modifies the breakdown of vertically propagating GWs with eastward phase speed in the summer mesosphere. More specifically, changes in the mean wind cause a shift in the breaking levels of these waves, which in turn modifies the equatorward residual circulation and, hence, upwelling and temperatures around the summer polar mesopause. A comprehensive demonstration of the causal chain of events giving rise to IHC is given by Karlsson et al. (2009) and by Körnich and Becker (2010).

GWs with phase speeds in the direction of and comparable to the mean flow are effectively absorbed. As a result, even though the troposphere radiates GWs with phase propagation into all horizontal directions, mainly westward (eastward) GWs will reach the winter (summer) mesosphere, where they break and deposit their westward (eastward) momentum (e.g., Lindzen 1981). Accordingly, a stronger and less variable eastward zonal wind in the winter middle atmosphere is more favorable for vertical propagation of westward GWs. Since the southern winter polar night jet is, on average, much stronger and less variable than its Northern Hemispheric counterpart, the net westward GW drag is stronger in the southern than in the northern winter mesosphere. Consequently, the residual circulation in the southern winter mesosphere is also stronger than in its northern counterpart, leading to a warmer winter polar stratopause (stronger downwelling) and a colder mesosphere in the equatorial region (stronger upwelling).

In the light of hemispheric differences, IHC has been suggested to play a role mainly by perturbing the Southern Hemispheric summer mesosphere as a result of the dynamically active northern winter stratosphere (e.g., Becker and Schmitz 2003; Karlsson et al. 2007; Körnich and Becker 2010). Hence, the approach to describe the signal propagation between the hemispheres is usually to consider dynamically active events in the winter stratosphere, which, via the mechanism summarized in the previous paragraph, result in a reduced strength of the residual circulation both in the summer and in the winter, leading to a warming of the summer polar mesopause region. Karlsson et al. (2009) showed that the lack of planetary wave activity in the winter polar stratosphere has a cooling effect on the temperature in the summer mesopause region, but little attention has been brought to this finding hitherto. We argue here that the seasonal state of the summer northern mesosphere is strongly tied to the winter Southern Hemisphere—and thereby “perturbed” by IHC—even to a larger extent than the summer southern mesosphere is affected, on average, by the winter Northern Hemisphere.

In Fig. 1, we show the zonal-mean temperature difference between July and January, averaged over the years between 2005 and 2012 from the Microwave Limb Sounder (MLS) instrument on the Aura satellite (e.g., Schwartz et al. 2008). In this figure, negative latitudes indicate the winter hemisphere. (We have constructed this figure by taking [mean July] minus [mean January] where the January field is flipped with respect to latitudes.) As can be seen, the high-latitude summer mesopause region is more than 5 K colder in July than in
January, which is to say that the NH summer mesopause is colder than the SH summer mesopause. We suggest that the temperature difference in this region could, to some extent, be a result of the IHC mechanism.

We hypothesize that the equatorial mesosphere is crucial for determining the seasonal zonal wind flow in the summer mesosphere and, hence, the breaking levels of GWs. In the equatorial region, the temperature is, as discussed previously, modified by the GW-driven branch of the residual circulation in the winter mesosphere. It is the strength of the winter-side flow that determines the strength of the adiabatic cooling in the equatorial mesosphere. It is evident in Fig. 1 that the winter mesosphere (latitudes between $-80^\circ$ and $-40^\circ$; yellowish area) is significantly warmer—in fact more than 20 K—in July than in January. This is a clear indication of the stronger July downwelling at high winter latitudes. Moreover, the equatorial mesosphere (marked with a dashed black oval in the figure) is significantly colder in July. It is this region that is of particular focus for our hypothesis.

In the absence of this winter-side equator-to-pole flow, the mesospheric equatorial region would be warmer. A warmer equatorial mesosphere would lead to lower breaking levels of GWs in the summer hemisphere and thus a warmer summer mesopause region, as seen by, for example, Karlsson et al. (2009). In the same fashion, an anomalously cold equatorial region leads to a cooling of the summer polar mesosphere (Karlsson et al. 2009; their Figs. 5–9).

An implication of the July–January differences in the summer mesospheric equator-to-pole temperature gradient in Fig. 1 is that the July summer zonal winds will be such that the GW breaking levels will be higher in the NH summer (July) mesosphere than in the SH summer (January) mesosphere (e.g., Becker and Fritts 2006; Karlsson et al. 2009). Thus, we argue that the negative temperature anomaly ($|\Delta T| > 5$ K) is denoted with a white line for clarification) in the summer mesopause region in Fig. 1 is, at least partly, attributable to higher GW breaking levels in the NH.

It should be pointed out that the January summer stratosphere is about 3–5 K warmer than the July summer stratosphere, as seen in Fig. 1, at latitudes between 40° and 80°. This difference has been discussed by, for example, Rosenlof (1996), Alexander and Rosenlof (1996), Siskind et al. (2003), and Becker et al. (2015), which all suggest that gravity waves with westward phase speed account for most of the hemispheric asymmetry in summer stratospheric temperature. As investigated by Becker et al. (2015), this temperature difference does not affect the filtering of gravity waves significantly after solstice (see their Figs. 6 and 11 and appurtenant discussion). However, part of the temperature differences seen in the summer mesosphere in Fig. 1 could be due to hemispheric differences in the tropospheric summer jet, as concluded by Siskind et al. (2003) and confirmed by Becker et al. (2015). Also, we cannot rule out differences in GW sources as an additional possible contributor.

Nevertheless, from the arguments presented above, we hypothesize that a middle atmosphere without orographic GWs in the winter hemisphere would result in an overall warmer summer polar mesopause region.

To test these arguments, we analyze sensitivity experiments with a new version of the Kühlungborn Mechanistic General Circulation Model (KCMC; Becker et al. 2015; Schlutow et al. 2014). In particular, we compare a control simulation with realistic representation of the dynamics of the middle atmosphere to special setups regarding the specification of gravity wave parameters in order to investigate how the summer mesosphere is affected by the residual flow in the winter mesosphere.

The remainder of this study is structured as follows: In section 2, we give a brief description of the KCMC. The model setup for this particular study is outlined in section 3, where our main results are also presented. Section 4 provides a discussion of our results, while the main findings are summarized in section 5.

2. Model description

The KCMC is a mechanistic GCM from the surface to the lower thermosphere (uppermost level around 0.000 03 hPa, corresponding to about 125-km height). It is employed here with a conventional resolution of T32 spectral truncation in the horizontal direction and 70 full layers in the vertical. The vertical discretization is in accordance with that of Simmons and Burridge (1981), and the vertical hybrid coordinate approaches pressure surfaces from the midtroposphere on. The model is equipped with an idealized radiation scheme, as well as with an explicit tropospheric moisture cycle. Land–sea contrasts are taken into account by including orography, land–sea masks for several parameters (relative humidity, heat capacity, and albedo), and a slab ocean with prescribed lateral heat flux convergence. These model components are strongly idealized but nevertheless allow for a simulation of a reasonable synoptic and planetary wave activity throughout the whole model domain. More details can be found in Becker et al. (2015).

Subgrid-scale parameterizations include a local boundary layer scheme in accordance with that of Holtslag and Boville (1993), a Smagorinsky-type horizontal diffusion [described by Becker and Burkhart (2007)] completed
by an enhanced linear diffusion in the lower thermosphere, and parameterizations of GWs. The latter consist of the McFarlane scheme for orographic GWs (McFarlane 1987) and an extended Doppler-spread parameterization for nonorographic GWs (Becker and McLandress 2009) that is based on the original ideas of Hines (1997a,b). In the KMCM, all subgrid-scale momentum fluxes are energetically closed by the corresponding thermal effects (i.e., frictional heating in the case of diffusion and the complete energy deposition in the case of GWs). Energy conservation is furthermore enforced by discretization of thermal effects, as proposed in Becker (2003). In summary, the KMCM is a climate model that is strongly idealized compared to comprehensive models like the Canadian Middle Atmosphere Model (CMAM), for instance (Fomichev et al. 2002), but is much more complex than dry models driven by temperature relaxation (e.g., Kushner and Polvani 2006).

Our model results are based on a control simulation where the model has been tuned with regard to climatology, wave–mean flow interaction, and GW effects in the middle atmosphere and lower thermosphere (MLT). This simulation—where the tuning has been somewhat improved compared to Becker et al. (2015)—includes the annual variation of the solar insolation owing to orbital eccentricity. As shown in the previous paper, eccentricity is only of minor importance for the simulated hemispheric differences in the middle atmosphere since these are mainly induced by the differences in tropospheric surface processes and the related forcing of planetary waves. As will be described in section 3, in addition to the 10-yr control simulation, we have performed two perturbation runs where the GW parameterizations (for nonorographic and orographic GWs) were set to zero in either the Northern or the Southern Hemisphere.

To better differentiate between these perturbations runs and the control run, we document here the main features of the control run with regard to climatology and wave drag. Figure 2 shows with colors the zonal-mean temperature $T$ and zonal wind $u$ for January and July. The inserted black contours correspond to the resolved PW drag (Figs. 2a,b) and the GW drag (Figs. 2c,d). The former is defined as the generalized Eliassen–Palm flux (EPF) divergence (in units of meters per second per day) due to zonal wavenumber 1–6 only. This includes the zonal drag owing to forced PWs (mainly in

![Figure 2: Zonal-mean climatology of the control simulation during (left) January and (right) July. Color shading shows (a),(b) temperature and (c),(d) zonal wind. The PW drag (shown above 900 hPa) and the GW drag are inserted as black contours for $1, 5, 10, 20, 40, 80,$ and $160 \text{ ms}^{-1} \text{ day}^{-1}$. See text for the definition of the PW drag and the GW drag.](image-url)
the northern winter stratosphere), in situ–generated PWs in both the summer and winter mesosphere, and the predominantly westward drag owing to thermal tides that are damped by horizontal diffusion in the thermosphere. The corresponding frictional heating gives a substantial contribution to the heat budget above the mesopause (not shown) that is ignored in other models. The displayed GW drag consists of the parameterized drag from nonorographic and orographic GWs, plus a minor contribution owing to resolved GWs (e.g., McLandress et al. 2006) that is approximated by the resolved EPF divergence minus the PW drag, as defined above.

The simulated wind and temperature fields are reasonable, although the winter upper mesosphere is generally too warm by about 10°–20°C, and the tropical cold point is somewhat too warm. The hemispheric differences due to the stronger PW drag in the northern winter stratosphere and lower mesosphere compared to that in southern winter are well captured (Figs. 2a,b). This is particularly evident from comparing the polar night jets in January and July (Figs. 2c,d). As a result, the wintery westward PW drag in the MLT is clearly stronger in July than in January below about 0.0003 hPa and even twice as strong around 0.03 hPa. From the discussion of the preceding section we expect a stronger winter residual circulation in the mesosphere during July than January, which is indeed confirmed by the model (not shown). Note also the stronger eastward GW drag and lower temperature around the summer polar mesopause in July in comparison to January.

The summerly westward PW drag centered around 50° latitude around 0.002 hPa is attributable to in situ–generated PWs (e.g., McLandress et al. 2006; Pendlebury 2012). This drag is significantly stronger in the Southern than in the Northern Hemisphere for the present model. This is consistent with a stronger vertical shear of the zonal wind at this latitude in the SH between about 0.05 and 0.005 hPa (not shown). This behavior agrees well with the observed nature of the 2-day wave, which is stronger in the southern than the northern summer mesopause region (e.g., Limpasuvan et al. 2000). Becker et al. (2015) showed that the stronger wind shear is triggered by higher polar temperatures in the lower mesosphere as a result of intrahemispheric coupling and orbital eccentricity and by higher equatorial temperatures around 0.03 hPa as a result of IHC. The summer polar mesopause is warmer in the SH than in the NH after solstice since the resulting eastward GW drag is weaker.

There is a positive PW drag in the southern winter polar mesosphere exceeding 20 m s⁻¹ day⁻¹. This feature is also simulated by the extended CMAM and was analyzed by McLandress et al. (2006). More specifically, the warmer polar night stratopause in July, which is induced by the stronger westward mesospheric GW drag, results in a baroclinic instability that induces eastward–traveling baroclinic planetary waves. Such a feature is usually not visible in the NH. Also note the significant westward PW drag in the lower thermosphere, which results from the damping of thermal tides and drives upper-level branches of the residual circulation toward the pole in either hemisphere (not shown).

3. Hypothesis testing

To test what effect the winter mesospheric residual circulation has on the summer mesosphere, we compare the control run described in section 2 to two perturbation runs. In these perturbation runs, the parameterized GWs were completely removed poleward of 20° latitude in either the Northern or the Southern Hemisphere (using tapering toward zero between the equator and 20° latitude for the GW amplitudes). Because we remove the parameterized GWs in the winter mesosphere, the winter hemisphere branch of the residual circulation is almost absent. Only resolved waves (including planetary waves, tides, and inertial GWs) remain and drive a weak poleward flow. This role of resolved waves cannot be excluded in our sensitivity experiment without introducing dynamically inconsistent damping mechanisms. Nor can we prevent the resolved waves and their interaction with the mean flow from differing among the control and perturbation runs. However, it is worth pointing out that the summer stratosphere remains the same in the control run and in the runs without GWs in the winter. Hence, we can rule out any anomalous GW filtering effects from lower down in the summer stratosphere (an intrahemispheric effect) that could otherwise affect our results.

Figure 3 shows the zonal-mean temperature and zonal-mean zonal wind differences, ΔT (Figs. 3a,b) and Δυ (Figs. 3c,d), respectively, between the control run and the perturbation runs: that is, (run with GWs in the winter) minus (run without GWs in the winter hemisphere) in colors. The warming of the respective winter stratosphere and lower mesosphere (Figs. 3a,b) due to GWs is evident. It is caused by the downwelling in the polar night region that is driven by the westward GW drag (see Figs. 2c,d), where orographic GWs contribute mainly to a warmer polar stratosphere. Also evident is the corresponding deceleration of the polar night jets (Figs. 3c,d). Since the nonorographic GW drag is much stronger in the southern than in the northern winter mesosphere, the warming around the polar night stratopause and the corresponding mesospheric
deceleration are also clearly stronger in July than in January.

There are additional effects on the MLT region, which are due to the removal of the winter GWs, as can be seen in Fig. 3. The areas with negative temperature response in the winter hemispheric MLT and the equatorial upper mesosphere result from the damping of thermal tides by GWs. That is, when including GWs in the winter hemisphere, tides become weaker, leading to cooling in the equatorial MLT due to reduced dissipative heating and cooling in the extratropics mainly due to weaker thermospheric poleward branches of the residual circulation. The reduced decelerating effect due to weaker tides in the control run is also visible as a generally positive zonal wind response in the lower thermosphere. As shown by Ortland and Alexander (2006), the damping of tides by parameterized GWs is mainly caused by the GW-induced vertical diffusion coefficient, which is on the order of 50 m² s⁻¹ in the northern winter mesosphere for the present model and even stronger in southern winter (not shown). The role of tides is further elaborated by Becker (2016, manuscript submitted to J. Atmos. Sci.).

To confirm that the responses we find are indeed caused by GWs and not by planetary waves, the experiments were repeated in a quasi-zonally symmetric model setup as far as the MLT is concerned. This is achieved by damping all zonal wavenumbers greater than zero above about 5 hPa such that planetary wave and tidal amplitudes become very small and the EPF divergence no longer contributes to the general circulation. Such a measure is easily implemented in a spectral GCM. This artificial model setup is comparable to the KMCM runs presented in Körnich and Becker (2010). Figure 4 shows the result from the quasi-zonally symmetric runs. When comparing Figs. 3 and 4, the same general picture arises from the two modeling approaches. In Figs. 4a and 4b, the cooling of the polar winter mesosphere due to GWs is mainly caused by the downward heat flux associated with the GW-induced
vertical diffusion (e.g., Liu 2000; Becker 2004). The extratropical wind response in the thermosphere in Figs. 3 and 4 exhibits a vertical extent of the deceleration in winter and the upward-shifted wind reversal in summer. Up to the mesopause, the model response in the quasi-zonally symmetric case is basically the same as for the regular KMCM case, in particular for the region around the summer mesopause. Hence, thermal tides and in situ–generated planetary waves do not significantly alter the overall effect of the coupling signal that is induced by GWs in the winter hemisphere. The main difference between the sensitivity experiments shown in Figs. 3 and 4 is that the overall GW effects on the mean flow are shifted to higher altitudes in the quasi-zonally symmetric case. This is consistent with Becker (2012), who showed that thermal tides, and presumably other traveling planetary waves as well, lead to a downward shift of GW breaking in the MLT.

Both Figs. 3 and 4 confirm that nonorographic GWs in the winter mesosphere indeed have a remote cooling effect in the summer polar mesopause region. Along with the equatorial cooling around 0.03 hPa, and thereby the induced zonal wind response at middle latitudes in the summer mesosphere between about 0.01 and 0.001 hPa, the pattern perfectly fits to the coupling mechanism proposed in section 1. Temperature differences just below the summer polar mesopause range from −2 to −16 K. The large negative values right above the summer polar mesopause (more than 100 K) are partly a result of a shift in mesopause heights between the control and perturbation runs and the rapid increase of temperature with height in the lower thermosphere. The temperature differences in the equatorial region are as low as −12 K for January (Figs. 3a, 4a) and about −25 K for July (Figs. 3b, 4b). In summary, the net effect of gravity waves in the winter middle atmosphere on the summer polar mesopause region is a substantial cooling.

As can be seen in Figs. 3, 4a, and 4b, the cooling signal in the summer mesosphere (due to GWs in the winter) is much stronger in July than in January. This asymmetry is attributed to hemispheric differences in the stratospheric winter flow, as discussed in section 1: The stronger eastward winter flow in the SH (July) gives rise to a stronger winter residual circulation driven by GWs than the one developing in the NH during January. A stronger upwelling and thereby a larger adiabatic cooling in the equatorial region during July than January results in a stronger enhancement of the summertime westward mesospheric flow, a more pronounced upward

![Fig. 4. As in Fig. 3, but for the quasi-zonally symmetric model setup (see text for details).](image-url)
shift in the GW drag (not shown), and thereby a stronger amplification of the summer mesospheric branch of the residual circulation.

Further evidence that what we see in Figs. 3 and 4 is an effect of the temperature change in the equatorial mesosphere is presented in Fig. 5. Here we show the zonal-mean temperature (Fig. 5a), the zonal-mean wind (Fig. 5b), and GW drag (Fig. 5c) averaged over specific latitude bands during July. Figure 5a shows the temperature averaged over 15°S–15°N. In the control run (black curve), where GWs are included in southern winter, the temperature between 1 and 0.01 hPa is on the order of 10 K lower than for the run without winter GWs (gray curve). In Fig. 5b, the mean zonal wind between 45° and 70°N, is shown. As a result of the modified meridional temperature gradient, the westward jet is weaker in the run without winter GWs (gray curve). In turn, the weaker mesospheric jet leads to lower GW breaking levels and to a reduction of the overall GW drag in the summer mesosphere. This is illustrated in Fig. 5c, which shows the GW drag averaged over the same latitude band for the two cases. Figure 5d shows the adiabatic heating in the polar mesosphere, between 70° and 90°N. In the case for which winter GWs are excluded, the adiabatic cooling is not as intense. Finally, the absolute temperatures over the mesospheric polar cap (again 70°–90°N) are shown in Fig. 5e. In accordance with our hypothesis, the summer polar mesosphere warms significantly in the absence of GWs in the winter hemisphere. The profiles for January (not shown) are very similar to the July ones presented in Fig. 5, but the responses are slightly weaker, as mentioned above.

For further confirmation that the IHC mechanism is suppressed when removing the winter GW, we examine how the variability in the middle atmospheric temperature field changes with respect to the winter stratosphere as the winter GW drag is removed. Figure 6 shows the covariance of the July and January monthly mean temperature fields with respect to the winter stratosphere [averaged over 100–10 hPa in altitude, 40°–60°S for July, and 60°–90°N for January, as discussed in Karlsson et al. (2007)] for the three different runs described previously: the control run (Figs. 6a,b), the run without winter GWs (Figs. 6c,d), and the run in which PWs are dampened (Figs. 6e,f). We choose to show the covariance field rather than the correlation since it adds information about the amplitude of the variability. The contoured areas in the figure, denoting regions of statistical significance, correspond roughly to correlation coefficients > 0.6. In Figs. 6a,b (the control runs), the characteristic pattern for IHC (see, e.g., Karlsson et al. 2009; their Fig. 2) shows up nicely, in particular for the July case (Fig. 6a). When the winter GWs are removed (Figs. 6c,d), the equatorial mesospheric variability is no longer significantly covarying.
FIG. 6. Covariance in the (left) July and (right) January temperature field with respect to the winter stratosphere (100–10 hPa; 40°–60°S for July and 60°–90°N for January) for (a),(b) the control run, (c),(d) the run without GWs in the winter hemisphere, and (e),(f) the run where planetary scale waves from below are damped at 5 hPa. White contoured regions correspond to correlation coefficients for which the $p$ value < 0.05 (a 5% significance level). Black [control run (solid) and runs without winter GWs (dashed)] contours denote the 140-K contour (the region of the summer mesopause) for each run.
with the winter stratosphere in July (Fig. 6c). Nevertheless, there is still a signal of downwelling in the winter mesosphere and also a tidal signature in the thermosphere. Note, however, that the removal of GWs in January (Fig. 6d) has less effect on the equatorial mesosphere. In fact, the quadruple structure in January is enhanced in the absence of winter GWs. We speculate that the remaining meridional flow may be a result of both the resolved GWs that are still able to make it up to the winter mesosphere and to PWs, which, in the absence of parameterized GWs might be able to propagate higher up in the atmosphere and interact with the mesospheric zonal flow in a similar fashion as the westward GWs otherwise do. Howbeit, the amplitude of the variability is reduced throughout the temperature field in July. In addition, the temperature reduction as a result of the removal of winter GWs can be traced by comparing the black contours denoting the summer mesopause (140 K; control runs are solid, and runs without GWs in winter are dashed) in Figs. 6a–d.

Figures 6e and 6f show the covariance of the monthly mean July and January temperature fields in the case where PW-scale features are heavily damped above 5 hPa. It is obvious from this figure that the IHC pattern is absent in the mesosphere. This supports our notion above that PWs (and other resolved waves) cause the variability pattern seen in Figs. 6c and 6d.

4. Discussion

The mechanism for the coupling discussed here—from the winter mesosphere to the summer mesosphere—is identical to the IHC mechanism described, for example, by Karlsson et al. (2009) and Körnich and Becker (2010). However, in contrast to previous studies, we focus on the role of IHC in a more fundamental sense; IHC has up until now mostly been associated with PW events. We show here that the coupling exists regardless of any stratospheric PW activity and that it is even more pronounced between the NH summer and the SH winter because of the stronger GW drag in the SH winter mesosphere. Hence, the seasonal mean state of the summer mesosphere in the NH is more affected by IHC than the corresponding mean state of the SH. The most important distinction between IHC as induced by stratospheric PW activity and the role of IHC presented in this study is that the former leads to anomalous warming and cooling events of the summer polar mesosphere, whereas the latter has a net cooling effect. Our sensitivity experiment qualitatively confirms the presence of an IHC that is constantly acting to cool down the summer polar mesosphere. In this sense, enhanced planetary wave activity in the winter stratosphere modulates the summer mesopause via a reduction of IHC relative to the mean state.

The simulated general circulation response in the MLT due to parameterized GWs in the winter hemisphere is qualitatively the same for both northern and southern winter, and hemispheric differences are mainly caused by hemispheric differences in the forcing of planetary Rossby waves (e.g., Becker et al. 2015). To investigate Earth’s eccentricity as a significant contributor to the hemispheric differences in the middle atmosphere, we repeated our simulation without orbital seasonality of the solar constant and found essentially the same model response for both July and January.

Another question concerns the relative importance of orographic and nonorographic GWs. Without showing the sensitivity experiment where only nonorographic GWs are eliminated in the winter hemisphere, we mention here that orographic GWs contribute to the residual circulation in a similar way as planetary Rossby waves: that is, they help to drive the residual circulation in the winter stratosphere up to about the stratopause. Therefore, removing only orographic GWs in the winter hemisphere leads to stronger nonorographic GW drag in the mesosphere and a colder summer mesopause, particularly for January. This is illustrated in Fig. 7, which shows the January control run minus a run where orographic GWs are absent. As seen, the orographic waves in the winter stratosphere contribute to a heating of the summer polar mesosphere.

Complementary to the KMCM results we also show that the July–January relative importance of IHC can be traced in MLS temperature data as well as in the nudged version of the extended Canadian Middle Atmosphere Model (CMAM30). The latter dataset is a version of the extended CMAM (Fomichev et al. 2002), nudged to ERA-Interim (see Simmons et al. 2007) data between the surface and 1 hPa. It is free running at higher altitudes and reaches up to about $10^{-6}$ hPa (approximately 210 km). Further information about the CMAM and CMAM30 can be found in, for example, de Grandpré et al. (2000), Scinocca et al. (2008), McLandress et al. (2014), Hegglín et al. (2014), and Shepherd et al. (2014). The data from the CMAM30 span the years between 1979 and 2009 and are provided by Environment Canada (CCCma 2013).

Figure 8 shows the correlation between the monthly mean temperature in the winter stratosphere (the time series spatially averaged over the area marked with a red box is our reference point) and the rest of the monthly mean temperature field, at each grid point, in July (Figs. 8a,c,e) and in January (Figs. 8b,d,f). In Figs. 8a and 8b, the July and January temperature correlation fields
retrieved from MLS data are shown for the years between 2005 and 2012. In Figs. 8c and 8d, we show the corresponding temperature correlation fields composed from CMAM30 data between 2005 and 2009: the years that overlap with the MLS dataset. The entire record of the CMAM30 temperature data between 1979 and 2009 is utilized in Figs. 8e and 8f. As can be seen by comparing Figs. 8a,c,e and 8b,d,f, the IHC pattern is more robust for July, both spatially and in terms of correlation strengths, than for January for all three rows of Fig. 8. This is in alignment with our arguments that the IHC mechanism plays a more important role for the seasonal state of the summer polar mesopause in July than in January: the variability superposed on the mean state will have a larger and a more robust impact on the NH summer mesosphere.

Complementary to the correlation fields depicted in Figs. 8e and 8f, we show in Fig. 9 the corresponding standard deviation (Fig. 9, top) and covariance fields (Fig. 9, middle). Not surprisingly, the standard deviation amplitudes in the winter polar stratosphere and mesosphere and also in the summer polar mesopause region (e.g., Bailey et al. 2007; Gumbel and Karlsson 2011) are larger in January than in July. In Fig. 9, middle, showing the covariance of the temperature field with respect to the winter stratosphere for July and January (i.e., the correlation fields of Figs. 8e and 8f multiplied with the standard deviation field in Fig. 9, top, respectively), regions of statistically significant correlation coefficients are contoured in white. Notice the resemblance with the KMCM results presented in Figs. 6a, 6b, and 6d: in both models, the robustness of the pattern is more pronounced in July (Figs. 6a and 9, middle left) than in January (Figs. 6b and 9, middle right). The same hemispheric characteristic correlation fields are also seen in Karlsson et al. (2009, their Fig. 1).

Finally, in Fig. 9, bottom, the first empirical orthogonal function (EOF), which indicates where the temperature field has its maximum variance [see, e.g., Hannachi et al. (2007) for information about the EOF technique] is shown. It is worth emphasizing that the EOF analysis is not carried out with respect to the winter stratosphere variability, as is the case for the results presented in Fig. 9, top and middle. Therefore, it is intriguing to see that the characteristic IHC pattern dominates the variance field in the middle atmosphere in both July and January.

It is also noted that the winter stratospheric temperatures correlate with the summer polar mesosphere at a higher altitude in the SH summer than in the NH summer. This could perhaps be explained by the stronger tropospheric jet in the SH summer: fewer GWs with eastward phase speeds reach the SH summer mesosphere (January). As a result, less eastward GW drag is available for redistribution by IHC, and this could be an additional reason for the SH summer being less affected by IHC, although the NH winter variance is strong.

5. Summary

We have investigated the influence that the GW-driven residual circulation in the winter hemisphere has on the summer polar mesopause region. Our findings are mainly based on sensitivity experiments using a mechanistic middle atmosphere climate model and can be summarized as follows.

IHC is fundamental in the sense that it is continuously affecting the summer mesosphere circulation as a result of the GW-driven residual circulation in the winter middle atmosphere. We therefore suggest an expansion of our view of the phenomenon: it is both a transient and a permanent feature.
The transient nature of IHC is that winter planetary wave activity occasionally reduces the net GW drag in the winter mesosphere, which then leads to a weaker winter residual circulation and a warmer summer polar mesosphere. The permanent effect of IHC is a cooling of the upper summer polar mesosphere that is determined by the strength of the westward GW drag in the winter mesosphere. This is complementary to what has hitherto...
been referred to as interhemispheric coupling, viewing it as an occasional perturbation rather than a continuous driver of summer mesosphere conditions.

IHC is crucial when it comes to tracking down mechanisms for middle atmospheric hemispheric differences. There are certainly other contributors, but to fully understand the particularly low temperatures of the summer mesopause in the Northern Hemisphere, the winter Southern Hemisphere appears to be decisive.

**FIG. 9.** CMAM30 monthly mean variability for (left) July and (right) January presented in (top) standard deviation, (middle) covariance with respect to the winter stratosphere (averaged over 100–10 hPa; 40°–60°S for July and 60°–90°N for January), and (bottom) the first EOF of the middle atmosphere. White contours in (middle) mark areas of the statistical significance level of 0.05 for the correlation coefficients (see Figs. 8e,f).
The individual steps described in the IHC mechanism are based on well-known concepts of wave–mean flow interactions; its characteristic signal is (statistically) present in models and in observations. Our sensitivity experiments qualitatively confirm the presence of an IHC that is constantly acting to cool the summer polar mesosphere. In this sense, the northern summer mesopause is much more affected by the coupling than the southern summer mesopause. This view explains in particular why the summer mesopause in July is significantly colder than the summer mesopause in January.

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