Climatological Effects of Orography and Land–Sea Heating Contrasts on the Gravity Wave–Driven Circulation of the Mesosphere

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ABSTRACT

On the basis of permanent January simulations performed with an idealized general circulation model for the troposphere and middle atmosphere, the sensibility of the general circulation to orographic and thermal forcing of large-scale stationary waves is assessed. Gravity waves are parameterized following Lindzen's saturation theory. Up to the stratopause, present model results coincide with earlier estimates, confirming that the boreal winter zonal-mean climate does crucially depend on the combined action of orography and land–sea heating contrasts. Since, in turn, the propagation and breakdown of internal gravity waves is strongly modulated by the background horizontal winds, the mesospheric response to stationary wave forcing turns out to be substantial as well. It is found that in the climatological zonal mean, a warmer polar night stratosphere is accompanied by lower temperatures in the mesosphere up to about 80 km. The temperature signal induced by stationary wave forcing changes sign again in the upper mesosphere/lower thermosphere, which, except for the polar night region, is globally heated up by 10–20 K. This heating is weaker if the assumed Prandtl number for gravity wave–induced vertical diffusion is raised from 3 to 6.

The thermal effects in the mesosphere are interpreted in terms of a global weakening of the summer-to-winter-pole residual circulation that occurs along with strongly diminished gravity wave drag, turbulent diffusion, and energy deposition in the northern winter mesosphere. The weakening of gravity wave effects in the presence of quasi-stationary planetary waves is dominated by reduced efficiency of gravity wave saturation in the mesosphere. That is, due to the more variable and, on average, reduced planetary-scale horizontal winds, gravity wave saturation is distributed over a greater depth and drops in altitude. On the other hand, enhanced critical level absorption of gravity waves in the lower stratosphere plays at most a secondary role. Furthermore, present model results suggest that the winter–summer asymmetry in gravity wave breakdown, which is well known from the northern mesosphere, may be absent or even reversed in the southern mesosphere.

1. Introduction

It is widely accepted that the stratospheric residual circulation is primarily driven by planetary waves (e.g., Rosenlof and Holton 1993; Holton et al. 1995; Becker and Schmitz 1999, hereafter BS1). Particularly in the lower stratosphere, the observed asymmetry between boreal and austral winter results from the strong orographic and thermal forcing of stationary waves in the boreal winter lower troposphere (Yulaeva et al. 1994). Above the stratopause, the zonal drag due to planetary waves diminishes, and the breakdown of internal gravity waves drives the summer-to-winter-pole residual circulation in the mesosphere (Lindzen 1981). Given this essential difference in the zonal wave drag between the lower and upper part of the middle atmosphere, one would not necessarily expect a strong influence of orography and land–sea heating contrasts on the upper mesosphere/lower thermosphere (MLT). For instance, Lübken et al. (1999) found no significant differences in the temperature profiles observed in the austral and boreal summer MLT. Nevertheless, the question of a possible north–south asymmetry in the MLT is discussed in the literature (e.g., Huaman and Balsley 1999; Dowdy et al. 2001). If such an asymmetry indeed exists, it is most likely that the hemispheric differences in the tropospheric forcing of planetary waves do play a major role.

An obvious influence of planetary wave activity on the mesosphere is well known from sudden stratospheric warmings. Such events are characterized by a breakdown of the polar vortex and substantial warming of the polar night lower stratosphere. In addition, a sudden stratospheric warming is accompanied by substantial cooling of the polar lower mesosphere (e.g., Labitzke 1972; Whiteway and Carswell 1994; Walterscheid et al. 2000). Corresponding internal variability patterns are also known from long-term GCM experiments including parameterization of gravity wave drag (Volodin and Schmitz 2001). Based on an idealized numerical model, Holton (1983) showed that, in the zonal mean, the cool-
ing of the mesosphere during a sudden warming can be explained by an almost complete critical level absorption (filtering) of internal gravity waves (IGWs) due to the reversal of the zonal-mean zonal wind in the stratosphere. As a result, the winter hemispheric IGW-driven branch of the residual circulation diminishes and cannot maintain mesospheric temperatures far above thermal equilibrium. This mechanism has been further investigated by Dunkerton and Butchart (1984) who showed that the longitudinal variations of the planetary-scale flow enable a significant portion of the gravity waves launched in the troposphere to pass through the stratosphere even during major warming events. Nevertheless, the effect found in Holton’s idealized experiment was in the main confirmed.

In the present study, we address the question whether a dynamical interaction between the stratosphere and the mesosphere, that is due to the modulation of gravity waves by planetary waves, is also possible on the climatological scale. Accordingly, we compare several long-term model simulations that differ with respect to planetary wave activity only. For convenience a simple general circulation model (SGCM) of the troposphere and middle atmosphere is employed. Gravity wave effects are incorporated following Lindzen (1981), using a family of 14 individual waves. The IGW momentum fluxes (or equivalently the energy fluxes) are launched around ~170 mb. The present IGW scheme is completed by the appropriate thermodynamic effects following Becker and Schmitz (2002, hereafter BS3). Since at present it is hardly known how the upper tropospheric excitation of IGWs depends in detail on the large-scale atmospheric flow, a fixed and horizontally uniform source spectrum is assumed. By this specification we focus on the modulation of gravity wave breakdown by planetary wave activity. A possible competing effect that may result from the modulation of IGW excitation is excluded. Other idealizations of the model correspond to representing radiative and latent heating rates by simple parameterizations, namely by temperature relaxation towards a zonally symmetric equilibrium temperature (Shine 1987) plus prescribed cumulus heating in the deep Tropics and self-induced condensational heating in the middle latitudes. The heating functions can independently be substituted by their zonal averages. Thus, the model idealizations allow to systematically assess the remote effects of orography, land–sea heating contrasts, and their combination on the general circulation of the middle atmosphere. As concerns the stratosphere, corresponding model results have confirmed the picture that synoptic and planetary waves must be considered as the primary cause of tropical upwelling and the stratospheric residual circulation (BS1). In addition, these numerical experiments have revealed a mutual amplification of orographic and thermal effects (Becker and Schmitz 2001, hereafter BS2). It is thus an important question whether and to what extent the planetary wave pump can affect the mesosphere by modulating the breakdown of IGWs on the climatological scale.

In the following section we will briefly describe the SGCM that has also been used in a companion study (BS3). Details of the gravity wave parameterization are given in the appendix. Section 3 summarizes the climatological changes that are generated by full forcing of stationary waves. Furthermore we present intermediate sensitivity experiments to demonstrate the mutual amplification of the remote signals owing to orographic and thermal forcing of stationary waves. In section 4, the mechanisms behind the effects in the mesosphere are explored in some detail. Our main results are summarized and discussed in the final section.

2. Model description

The SGCM and its climatology are described in more detail elsewhere (BS2 and BS3). In the following we focus on those aspects that are important in the context of the present study.

Whereas the dynamical core of the model is GCM standard (60 hybrid levels extending from 990 up to 0.0003 mb and T29 spectral resolution), the main model physics is comprised in the idealized representation of radiative and latent heating. In particular, we employ temperature relaxation, prescribed convective heating in the deep Tropics (e.g., Hou 1993), and self-induced condensational heating in the midlatitudes (Mak 1994). Then the complete diabatic heating in terms of temperature tendency is defined as

\[ \text{diabatic heating} = -\frac{T - T_E}{\tau} + Q_c + \frac{\|\mathbf{h}(\mathbf{u})\|}{40 \text{ mb d}^{-1}} Q_m + \text{diffusion} + \text{dissipation}. \]  

Here, \( T \) is temperature, and \( T_E \) is equilibrium temperature (see Fig. 3a in BS3). The relaxation time \( \tau \) is horizontally uniform. It amounts to 17 days in the troposphere, increases to 50 days around the 100-mb level, and slopes down to 7 days in the upper stratosphere and mesosphere. It differs from Fig. 1b in BS3 by including a considerably long relaxation time in the lower stratosphere. Such a feature is known from radiation model calculations (e.g., Gille and Lyjak 1986) and it has also been accounted for by Dunkerton (1991) within an idealized model. We have found that this modification leads to somewhat enhanced transient planetary wave activity in the middle atmosphere while the overall model climatology is not changed significantly.

Self-induced heating [third term on the rhs of (1)] is scaled by the pressure velocity \( \mathbf{u} \) and vanishes in the case of descent as indicated by the Heaviside function \( \mathbf{h}(\mathbf{u}) \). The horizontal structures of \( Q_c \) and \( Q_m \) represent the tropical convection zones and land–sea heating contrasts in the northern winter troposphere, respectively (see Fig. 2 in BS2). The overall thermal forcing owing to \( T_E, Q_c, \) and \( Q_m \) describes perpetual January condi-
tions. Both heating functions $Q_s$ and $Q_{m}$, as well as orography $\Phi_s$, can independently be substituted by their zonal averages $\langle Q_s \rangle$, $\langle Q_{m} \rangle$, and $\langle \Phi_s \rangle$.

The vertical diffusion term on the rhs of (1) includes vertical diffusion of potential temperature due to boundary layer turbulence (Prandtl number 1), horizontal diffusion of temperature (Prandtl number 2), and IGW-induced vertical diffusion. For the latter diffusion tendency, a Prandtl number of 3 is employed, which fits within the range of existing estimates (e.g., Hocking 1999; Thomas 1996). The model sensitivity to the choice of this parameter is discussed in section 4c.

For convenience, we will refer to dynamic heating as the sum of advection and adiabatic heating by the resolved planetary-scale flow. In the middle atmosphere, this heating is to a good approximation equal to the adiabatic heating associated with the residual circulation. Note that energy deposition owing to IGW saturation represents dynamic heating as well. In section 4 we will nevertheless consider the sum of heat diffusion, frictional heating, and energy deposition in order to sum up the direct thermodynamic effects owing to IGW saturation, which will be referred to as direct IGW heating. For the sake of completeness, this definition includes also the contributions from horizontal diffusion.

Our gravity wave parameterization is based on the saturation theory of Lindzen (1981). A fixed and horizontally uniform source spectrum consisting of 14 individual waves in eight equidistant azimuths is launched around 170 mb. The waves’ momentum fluxes are precisely prescribed by the parameter setting such that an intercomparison of model climatologies with different stationary wave forcing configurations does not depend on arbitrarily different gravity wave sources. Details of the present implementation of the Lindzen scheme are given in the appendix.

3. Remote mesospheric effects induced by stationary wave forcing in the troposphere

In the following we inspect the climatologies of two long-term integrations. In the first model run (1200-day time series) we apply full forcing of stationary waves owing to orography, self-induced midlatitude heating, and tropical heating. Such a run has also been analyzed in BS3. The second model run (180-day time series) employs the zonal means of orography $\Phi_s$ and heating functions only; it thus represents the equivalent rotationally invariant simulation, which will also be referred to as aquaplanet run.\(^1\) As in BS1, these experiments are abbreviated as $\Phi_s(\lambda)/Q_s(\lambda)/Q_s(\lambda)$ and $\Phi_s(\lambda)/Q_s(\lambda)/Q_s(\lambda)$ in order to define which field $\Phi_s$, $Q_s$, or $Q_m$ is applied with full longitude dependence or as zonal mean only. This notation is also suitable to define intermediate runs.

Figure 1 displays zonal-mean temperature and zonal wind for both the full wave forcing and the aquaplanet simulation. The black contours in Fig. 2 show the corresponding residual mass streamfunctions and Eliassen–Palm flux (EPF) divergences, which were calculated in quasigeostrophic approximation following Edmon et al. (1980). In addition, the residual meridional wind, as well as the averaged zonal wind tendency generated by IGWs via momentum deposition and vertical momentum diffusion, are indicated by white contours in the upper and lower panels of Fig. 2. In order to allow for a quantitative comparison of the so-defined zonal wave drag due to planetary waves and gravity waves, either quantity is computed as zonal wind tendency and plotted in units of meters per second per day.

Comparison of the lhs and rhs panels in Figs. 1 and 2 clearly confirms the established picture of the stratospheric residual circulation (e.g., Holton et al. 1995; BS1): tropical upwelling, as well as the poleward stratospheric mass flux, are substantially amplified as a result of the planetary wave drag that is induced by the combined action of orographic and thermal forcing of stationary waves; the corresponding amplifications in adiabatic heating/cooling and zonal deceleration are reflected by the changes in zonal-mean temperature and zonal wind.

As concerns the thermal structure of the mesosphere, the situation is more complex. The differences in Fig. 1b from Fig. 1a (see also Fig. 3f) indicate cooling of the lower winter mesosphere while the upper mesosphere outside the polar night region is globally heated up by about 10–20 K. Furthermore, the winter mesospheric residual circulation is strongly reduced as indicated by the streamfunction contours in Figs. 2a and 2b. The effect shows up more clearly as a reduction of the Northern Hemisphere maximum of the residual meridional wind from 22 to 12 m s\(^{-1}\). At first sight, this is surprising since in run $\Phi_s(\lambda)/Q_s(\lambda)/Q_s(\lambda)$, a strong zonal deceleration of more than $-10$ m s\(^{-1}\) day\(^{-1}\), owing to EPF divergence, occurs around 0.02 mb (Fig. 2d). However, at these altitudes, the residual circulation is primarily driven by internal gravity waves. The maximum gravity wave drag shrinks from $-210$ to about $-110$ m s\(^{-1}\) day\(^{-1}\) from run $\Phi_s(\lambda)/Q_s(\lambda)/Q_s(\lambda)$ to $\Phi_s(\lambda)/Q_s(\lambda)/Q_s(\lambda)$. As a result, the overall maximum wave drag is reduced by about a factor of 2 consistent with the upper level weakening of the residual meridional wind.

We now consider the separate model sensitivities to orographic and thermal forcing of stationary waves. Similar to the findings of Hamilton (1995), the present model shows no significant signal in the stratosphere and mesosphere in response to imposed changes of diabatic heating in the deep Tropics. Therefore, the middle atmospheric model response to thermal stationary wave forcing is in the main due to longitude-dependent self-induced heating. Figures 3a,b and 3c,d show the climatological differences of zonal-mean zonal wind and

\(^1\) The zonal means of $\Phi_s$ and $Q_s$ formally deviate from aquaplanet conditions. However, their effects are hardly visible in the climatology of the aquaplanet simulation.
temperature in the intermediate runs from the aquaplanet run. The intermediate runs are abbreviated as \( \Phi_s(\lambda)/[Q_m]/[Q_c] \) or \( [\Phi_s]/Q_m(\lambda)/Q_c(\lambda) \) and employ either orographic or thermal forcing of stationary waves only. For comparison, Figs. 3e,f show the wind and temperature signals generated by the combined action of orographic and thermal stationary wave forcing, that is, the differences in Figs. 1d and 1b from Figs. 1c and 1a, respec-
Fig. 2. (a), (c) Residual circulation in the aquaplanet run and in (b), (d) the full stationary wave forcing experiment. (a), (b) The residual mass streamfunction is drawn with black contours for 0.01, 0.1, ±1, ±2, ±4, ±10, 50, 100, and 150 × 10⁸ kg s⁻¹. The residual meridional wind is indicated by white contours for 5, 10, 15, and 20 m s⁻¹. (c), (d) The Eliassen-Palm flux divergence (black contours) is calculated in quasigeostrophic approximation following Edmon et al. (1980). It is plotted as tendency of zonal wind for ±2, ±4, ±8, and 16 m s⁻¹ day⁻¹. In addition, the white contours give the total zonal drag (plotted for ±50, ±100, −150, and −200 m s⁻¹ day⁻¹) owing to momentum deposition and vertical momentum diffusion generated by IGW saturation.
Fig. 3. Climatological zonal-mean changes of zonal wind and temperature in the middle atmosphere induced by stationary wave forcing in the troposphere. (a),(b) Run $\Phi_1(\lambda)/[Q_m]/[Q_c]$ with orography as stationary wave forcing only minus aquaplanet run $\Phi_1/[Q_m]/[Q_c]$. (c),(d) Run $\Phi_1/[Q_m]/[Q_c]$ with thermal stationary wave forcing only minus run $\Phi_1/[Q_m]/[Q_c]$. (e),(f) Run $\Phi_1(\lambda)/Q_m(\lambda)/Q_c(\lambda)$ with full stationary wave forcing minus run $\Phi_1/[Q_m]/[Q_c]$. In (a),(c), and (e) the contour interval is 10 m s$^{-1}$, while it is 5 K in (b),(d), and (f). Zero contours are not drawn, and negative values are shaded.
The wind and temperature signals displayed in Fig. 3 bear roughly the same structure but differ strongly in amplitude. The nonlinear model response to full stationary wave forcing is generally much stronger than in the intermediate cases. The mutual amplification of the effects of orography and land–sea heating contrasts on the general circulation of the stratosphere (BS1 and BS2) is reproduced. In addition, present model experiments indicate that this mutual amplification is efficient also in the mesosphere.

Inspection of Fig. 2 has suggested that the climatological signals in the MLT visible in Figs. 1 and 3 are associated with changes in gravity wave effects. This conclusion is further confirmed if we repeat our sensitivity experiments with all gravity wave effects (momentum deposition, energy deposition, and IGW-induced vertical diffusion) turned off. Figure 4 shows corresponding climatological effects analogously to Figs. 3e,f. Above the stratopause, the zonal wind and temperature signals in Figs. 3e,f and 4a,b differ qualitatively from each other. Especially the strong cooling of the lower winter mesosphere and the warming of the MLT are absent if gravity waves are not accounted for.

Summarizing, IGW effects in the mesosphere strongly depend on stationary wave forcing in the troposphere. We expect that this dynamic link results from the modulation of gravity wave breakdown via changes of the background wind in the stratosphere and mesosphere. In the following we inspect this mechanism in some more detail.

4. Modulation of gravity waves by planetary waves

a. Winter hemisphere

Figures 5a,b display dynamic heating (adiabatic heating plus advection by the resolved planetary-scale flow) for both the full stationary wave forcing experiment \( \Phi_s(\lambda)/Q_{m}(\lambda)/Q_{c}(\lambda) \) and the equivalent aquaplanet run \([\Phi_s]/[Q_m]/[Q_c]\). Contours are drawn for 1, 2, and 3 K day\(^{-1}\) below 3 mb, and for \( \pm 5, \pm 10, \) and \( \pm 20 \) K day\(^{-1}\) at greater heights. In the polar night stratosphere between 150 and 3 mb, we observe a strong increase in dynamic heating from run \([\Phi_s]/[Q_m]/[Q_c]\) to run \( \Phi_s(\lambda)/Q_{m}(\lambda)/Q_{c}(\lambda) \), consistent with the strengthening of the residual circulation discussed in the previous section. Stratospheric changes in dynamic heating are generally balanced by associated changes in radiative heating, which can be described approximately by temperature relaxation (Yulaeva et al. 1994). Therefore, one expects that the stratospheric dynamic heating signal induced by stationary wave forcing is reflected by the temperature signal. This relationship is consistent with Fig. 3f.

A tight connection between the changes in dynamic heating and temperature is also valid in the winter mesosphere up to about 0.01 mb. In this altitude range, the remote effect of tropospheric stationary wave forcing contrasts with the enhanced planetary wave EPF divergence. However, here the changes in dynamic heating and temperature rather reflect the reduced mesospheric gravity wave drag (white contours in Figs. 2c,d) that results in diminished descent.

Substantial changes in IGW saturation are also evident from Figs. 5c,d which show the direct IGW heating for both model runs \([\Phi_s]/[Q_m]/[Q_c]\) and \( \Phi_s(\lambda)/Q_{m}(\lambda)/Q_{c}(\lambda) \). In section 2, this contribution to the heat budget has been defined as the sum of energy deposition by gravity waves and turbulent heating. In the present model, turbulent heating is dominated by IGW-induced vertical diffusion of potential temperature, which cools the MLT. Figures 5c and 5d indicate that this cooling even exceeds the energy deposition, which has maximum heating rates of more than 10 K day\(^{-1}\) (not shown). As
a consequence of stationary wave forcing in the troposphere, the (negative) direct IGW heating in the winter extratropical mesosphere is efficiently reduced. Thus, there is a strong cancellation between the changes in dynamic heating and direct IGW heating. As a result, above 0.1 mb or so, the temperature signal no longer corresponds to the changes in dynamic heating. Rather, it reflects the residuum of the changes in dynamic heating and direct IGW heating. Below 0.01 mb, the dynamic heating signal is generally dominant. At higher altitudes and between 30°N and 60°N, the temperature signal is positive (see Fig. 3f), indicating that the IGW heating signal overcompensates the changes in dynamic heating. Farther poleward, the upper mesospheric temperature changes are considerably weak. Note also that an approximate compensation between warming in the extratropics and cooling in the Tropics (or vice versa), that is well known from the stratosphere (Yulaeva et al. 1994), ceases in the upper mesosphere (Fig. 3f). This finding is again consistent with the strong direct IGW heating in this altitude range.

Why are the simulated gravity wave effects in the northern mesosphere diminished in presence of planetary waves? In Fig. 6 we reconsider the differences in gravity wave breakdown between the aquaplanet run and the full stationary wave forcing experiment. Figures 6a,b show latitude–height cross sections of the time and zonal-mean vertical flux of zonal momentum owing to our Lindzen-type gravity wave parameterization. Alternatively, we consider in Figs. 6c,d only those individual gravity waves of the source spectrum that become saturated before encountering a critical level. Only these “unstable” gravity waves contribute to the turbulent diffusion coefficient which is shown in Figs. 6e,f.

In the winter lower stratosphere, the westward IGW momentum flux is significantly reduced as a result of stationary wave forcing (Figs. 6a,b). This effect may be attributed to enhanced filtering or saturation.
Fig. 6. (a),(c),(e) Long-term zonal-mean gravity wave effects in the aquaplanet run and (b),(d),(f) in the full stationary wave forcing experiment. (a),(b) Vertical flux of westerly momentum (contour interval 0.4 g m$^{-2}$ s$^{-2}$). (c),(d) Same as (a),(b), but disregarding those individual gravity waves that propagate conservatively until encountering a critical level. (e),(f) IGW-induced vertical diffusion coefficient (contours 100, 300, 500, 700, ... m$^2$ s$^{-1}$). Zero contours are not drawn and negative values are shaded.
Comparing Fig. 6a to 6c or Fig. 6b to 6d, we can infer that gravity waves, which propagate conservatively until they encounter a critical level, do not contribute to the IGW momentum flux above 10 mb or so. Hence, above that level, IGW breakdown is dominated by saturation. Furthermore, we see that in run $\Phi_s(\lambda)/Q_m(\lambda)/Q_c(\lambda)$, the mesospheric momentum deposition is located at lower altitudes and that it is efficiently smoothed out. This behavior can be interpreted on the basis of the dispersion relation for gravity waves [Eq. (A7)] in association with the saturation hypothesis: for a gravity wave with easterly phase velocity, the altitude where saturation sets in—that is, the breaking level—decreases in presence of a weaker westerly flow. Consequently, temporal variability of the background wind causes a deeper distribution of breaking levels. A reduced mean vertical wind shear additionally increases the depth of saturation. By these mechanisms, the statistical changes of the winter mesospheric planetary-scale winds can reduce the efficiency of gravity wave saturation. Planetary wave–induced reductions of the mean zonal wind and its vertical shear are evident from Figs. 1 and 3. Figure 7 shows corresponding standard deviations of the zonal-mean zonal wind. Throughout the middle atmosphere, poleward of 40°N, there is essentially no internal variability in the aquaplanet run. On the other hand, the standard deviation in the simulation with full forcing of stationary wave is quantitatively comparable with observational estimates for January (Randel 1992).

Summarizing, the diminished gravity wave effects in the extratropical winter mesosphere are due, first, to enhanced filtering or saturation in the stratosphere, and second, to reduced saturation efficiency in the mesosphere itself. In order to assess the relative importance of both mechanisms, let us consider the individual wave momentum fluxes owing to those gravity waves that have easterly phase velocity components and hence are important for gravity wave breakdown in the winter middle atmosphere. Figure 8 shows these fluxes for the 10 mb layer in the long-term zonal mean. The azimuths of propagation are indicated in the title of each panel. Each momentum flux is due to two individual waves that differ with respect to their phase speeds but have identical wave parameters otherwise. For further reference see the appendix and Fig. A1. In the full stationary wave forcing run (upper dotted lines), the vertical momentum fluxes are reduced at most by about 30% compared to the aquaplanet simulation (upper solid lines). Particularly around 50°N, the reduction is not in excess of 20%. On the other hand, IGW momentum deposition (Fig. 2), direct IGW heating (Fig. 5), and IGW-induced turbulent diffusion coefficient (Fig. 6) are reduced by factors of 2–4 in the midlatitude winter mesosphere. Therefore, the diminished saturation efficiency in the mesosphere clearly dominates the simulated climatological effects of orography and land–sea heating contrasts on the winter mesosphere.

The lower lines in Fig. 8 show those portions of the respective momentum fluxes that are removed from the gravity wave source by critical level absorption only. The representation reveals that in the winter middle atmosphere, the filtering of IGWs with easterly phase velocity components is negligible for aquaplanet condi-
tions (lower solid lines). In the case of full stationary wave forcing (lower dotted lines), the enhanced filtering can at most partly explain the momentum flux reduction at 10 mb. Therefore, both enhanced gravity wave filtering and enhanced saturation appear to be roughly of equal importance in the stratosphere.

b. Summer hemisphere

Figure 3f has indicated that the thermal structure around the summer mesopause is sensitive to orographic and thermal forcing of stationary waves in the winter troposphere. Comparison of Figs. 3f and 4b proves that this sensitivity must somehow be related to the modulation of gravity wave effects. However, south of 30°N, the climatological changes in dynamic heating and direct IGW heating are rather weak. Alternatively, Fig. 9 shows the difference of the residual meridional wind in the full stationary wave forcing experiment from the aquaplanet run. There is a global reduction of the upper-level residual circulation. Hence, the associated dynamic heating signal (not shown) dominates the temperature signal in the upper mesosphere south of about 30°N.

c. Sensitivity to model parameters

In accordance with our previous SGCM studies (BS1, BS2), it has been found that the computational results presented here are generally robust against changes of the equilibrium temperature (see Fig. 3a in BS3), heating functions, resolution, and diffusion parameters. The mesospheric effects may nevertheless be sensitive to the assumed gravity wave parameter setting. As mentioned in section 2 and in the appendix, we use a modified gravity wave source compared to BS3. Nonetheless, the simulated climatological changes due to stationary wave forcing has been found to be quite robust even against this modification of the model.

The present IGW parameters have been adjusted in order to yield a reasonable climatology in the mesosphere, including the reversal of the upper-level zonal wind and the IGW-induced cold summer mesopause. In this respect, the specification of the Prandtl number, which is particularly crucial to the representation of the direct gravity wave heating, is more arbitrary. As shown in our companion study (BS3), the model climatology in the MLT undergoes significant changes if the Prandtl number is changed by a factor of 2 or so. It is thus worthwhile to ask how the modulation of gravity wave
effects is affected by such a parameter variation. Accordingly, we have repeated the full stationary wave forcing experiment $\Phi_s(\lambda)/Q_m(\lambda)/Q_c(\lambda)$ and the equivalent aquaplanet run $[\Phi_s]/[Q_m]/[Q_c]$, employing a Prandtl number of 6 instead of 3 for vertical diffusion of potential temperature in the MLT. Figure 10 shows the climatological zonal-mean differences between both runs with regard to zonal wind and temperature. Comparison with Figs. 3e,f indicates that the effects discussed so far apply, at least qualitatively, also in the case with a Prandtl number of 6. An exception is visible in the upper mesosphere north of about 60°N only. Here the temperature signal is now clearly negative instead of being indifferent.

5. Summary and concluding discussion

A condensed set of idealized numerical experiments has been investigated in order to assess the role of orography and land–sea heating contrasts in the boreal winter climatology of the mesosphere. Our main results may be summarized as follows.

The long-term changes owing to stationary wave forcing are characterized by a more than 20-K-warmer polar night stratosphere and lower temperatures in the polar night lower mesosphere (Fig. 3f). These signals are approximately compensated by reverse temperature changes in the Tropics and subtropics in agreement with Yulaeva et al. (1994). Up to altitudes of about 80 km (0.01 mb), the simulated temperature signal is due to the changes in the zonal wave drag that drives the residual circulation. In the stratosphere, the planetary wave pump is enhanced by stationary wave forcing (Holton et al. 1995; BSI), while the weakening of the residual circulation in the mesosphere is related to diminished zonal forcing owing to gravity wave breakdown.

The climatological changes of the residual circulation and the associated temperature signal in the winter middle atmosphere are reminiscent of sudden stratospheric warming events (e.g., Labitzke 1972). During a major warming, gravity waves are strongly filtered at critical levels in regimes of stratospheric easterlies and hence do not drive a strong residual circulation in the winter MLT (Holton 1983; Dunkerton and Butchart 1984). A distinct interpretation applies in the present long-term simulations. Stationary wave forcing generates a strong internal variability, particularly in middle and polar latitudes, which is accompanied by a reduced mean polar night jet (Figs. 7 and 1). As a result, the averaged saturation heights (breaking levels) are shifted to lower altitudes, and gravity wave–mean flow interaction is distributed over a greater depth. This leads to a considerable reduction of the efficiency of gravity wave saturation. In particular, the averaged maximum momentum deposition and direct gravity wave heating per unit mass or the IGW-induced turbulent diffusion coefficient are reduced by factors of 2–4 in the midlatitude winter mesosphere (Figs. 2, 5, and 6). On the other hand, enhanced critical level absorption or enhanced saturation in the lower stratosphere can explain at most a minor fraction of these effects (Fig. 8). In this context it should be noted that only two major stratospheric warmings with zonal-mean easterlies in the polar night stratosphere up to $-30 \text{ m s}^{-1}$ are generated in the 1200-day simulation with full forcing of stationary waves. It is therefore possible that in more realistic simulations frequent major warming events may give an additional contribution to the long-term reduction of gravity wave effects via enhanced critical level absorption. Never-
theless, we expect that the reduction of saturation efficiency will be the dominant process.

A further remote effect of orography and land–sea heating contrasts consists of a warming of the upper mesosphere/lower thermosphere by about 10–20 K at all latitudes south of about 60°N. Between 30°N and 60°N, this temperature signal is dominated by reduced cooling due to vertical diffusion. Conversely, in the tropical and summer hemispheric upper mesosphere, the positive temperature signal reflects reduced adiabatic cooling by the meridional circulation. Thus, the reduction of gravity wave breakdown in the extratropical winter mesosphere is accompanied by a global reduction of the gravity wave–driven branch of the residual circulation (Fig. 9). This sensitivity of the upper-level residual circulation to orography and land–sea heating contrasts is reminiscent of the situation in the stratosphere, however with opposite sign. The overall mesospheric signals also give a hint of how intraseasonal and interannual variability in the winter stratosphere can force global variability patterns up to the mesopause.

The long-term changes of zonal-mean temperature and zonal wind, that in the model are induced by the inclusion of orography and land–sea heating contrasts, are roughly consistent with the observed asymmetry between boreal and austral winter in several respects. This can be demonstrated for instance by plotting the climatological differences between January and July on the basis of Committee on Space Research (COSPAR) International Reference Atmosphere (CIRA) 86 (Fleming et al. 1990) and Met Office data (SPARC Data Center 2000), using a reversed latitude for July. Fig. 11 shows corresponding representations. The positive temperature differences in the summer stratosphere reflect enhanced ozone heating associated with the shorter distance between the earth and the sun during the boreal winter season. Apart from this feature, the signals shown in Fig. 11 may be compared to Figs. 3e,f or Figs. 10a,b.
For instance, Figs. 11a,b indicate a warmer winter stratosphere and a colder winter mesosphere during January, consistent with Figs. 3f and 10b. Furthermore, a warmer upper mesosphere is suggested by Fig. 11a, as found in the simulations. Also the zonal-mean zonal wind differences in Figs. 11c and 11d are roughly consistent with Fig. 3e and 10a. Thus, in view of the simplicity of the applied SGCM, the full stationary wave forcing experiment and the equivalent aquaplanet simulation may be considered to approximately represent the middle atmospheric general circulation during boreal and austral winter, respectively.

This latter conclusion motivates further comparison of model results with observational analyses. For instance, a comparison of Figs. 1a and 1b suggests that the austral summer mesopause [run $\Phi_s/(Q_{o0}/Q_s(\lambda))$] should be a few degrees warmer than its boreal summer counterpart [run $\Phi_s/(Q_{o0}/Q_s(\lambda))$]. Such a conclusion has been drawn by Huaman and Balsley (1999) based on a collection of several available datasets. These authors also found that around the summer mesopause, the westerly zonal wind component is stronger during austral summer than during boreal summer. This is in agreement with Figs. 3e, 10a, and 11c. Finally, as documented by Huaman and Balsley (1999) and Dowdy et al. (2001), the equatorward wind component in the region of the summer mesopause is weaker in January than in July. Obviously, this finding is consistent with a reduction of the summer-to-winter-pole residual circulation due to stationary wave forcing as found in the present model experiments (Fig. 9). Given these consistencies concerning north–south asymmetries in the summer mesosphere, one might also speculate on north–south asymmetries in the winter mesosphere. According to Figs. 5 and 6, gravity wave effects and turbulent parameters should be much stronger in austral winter (run $\Phi_s/(Q_{o0}/Q_s(\lambda))$) than in boreal winter (run $\Phi_s/(Q_{o0}/Q_s(\lambda))$). Hence, their strong asymmetry between summer and winter as observed in the Northern Hemisphere (e.g., Lübken 1997) may be absent or even reversed in the Southern Hemisphere. Again this could be understood as a remote effect of orography and land–sea heating contrasts.

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APPENDIX

Details of the Gravity Wave Parameterization

The vertical flux of horizontal momentum owing to IGWs, the associated pressure flux (or sensible heat flux), and the wave-induced turbulent diffusion coefficient may be written as

$$F := j_0^{-1} \sum_{j=1}^{N} F_t e_{\alpha_j}$$

with

$$e_{\alpha_j} := \cos \alpha_j e_x + \sin \alpha_j e_y,$$

(A1)

$$F_p = -j_0^{-1} \sum_{j=1}^{N} (\Pi_j - c_j) F_t,$$

(A2)

$$\vartheta := j_0^{-1} \sum_{j=1}^{N} D_j,$$

(A3)

Here, $j_0$ is the total number of individual waves. The $\alpha_j$ denote the respective azimuths of propagation, and the corresponding horizontal phase speeds are $c_j > 0$. The unit vectors in zonal and meridional direction are abbreviated as $e_x$ and $e_y$. Then $j_0^{-1} F_t e_{\alpha_j}$ and $j_0^{-1} D_j$ are the individual wave contribution to the momentum flux and diffusion coefficient. Furthermore, we define

$$\Pi_j := v \cdot e_{\alpha_j},$$

(A4)

where $v$ is the planetary-scale horizontal wind simulated by the model.

The vertical discretization of the SGCM is based on the method proposed by Simmons and Burridge (1981). For this scheme it is appropriate to specify $\mathbf{F}$ on intermediate half-model layers. The initial half level $z_{\alpha_j}$ is located around 170 mb. Writing the height argument in brackets and following otherwise the notation of Holton (1982), we have

$$F_{\alpha_j} := F_j[z_{\alpha_j}] = k_j \rho \left[ z_{\alpha_j} \right] \left[ H \right]^{-1} \exp \left[ \frac{z_{\alpha_j}}{H} \right] > 0.$$  (A5)

Here, $N$ and $\rho$ denote the Brunt–Väisälä frequency and background density of the model. The velocities $\bar{\alpha}_j > 0$ and the horizontal wavenumbers $k_j > 0$ are gravity wave parameters to be specified. The scale height $H = 7.5 \text{ km}$ is an additional parameter chosen as in BS3.

Critical levels $z_{\alpha_j}$ are implicitly defined by

$$\Pi_j - c_j = 0 \quad \text{at} \quad z = z_{\alpha_j},$$

(A6)

above which we generally have $F_j = 0$. Since the gravity waves’ vertical wavenumbers

$$m_j = N(\Pi_j - c_j),$$

(A7)

must be negative for downward phase propagation (or upward pressure flux), individual gravity waves with $\Pi_j - c_j > 0$ are also removed from the spectrum. If the condition $\Pi_j - c_j \geq 0$ is not fulfilled somewhere above the initial level, $z_{\alpha_j}$ is identified with the half layer between the uppermost two full-model layers. Below the critical levels, the individual gravity waves may reach static instability at the breaking levels $z_{\alpha_j}$, implicitly defined via

$$3H \ln \left[ \left( c_j - \Pi_j / \bar{\alpha}_j \right) \right] - z = 0 \quad \text{at} \quad z = z_{\alpha_j}.$$  (A8)

If the lhs of (A8) does not change sign between $z_0$ and $z_{\alpha_j}$, we formally set $z_{\alpha_j} = z_{\alpha_j}$.

Between $z_{\alpha_j}$ and $z_{\alpha_{j+1}}$, the waves are damped such as to
maintain marginal static instability. In Lindzen's theory, this determines the turbulent diffusion coefficient as well as the wave momentum fluxes. The latter are given by

\[
F_{bj} := pk(2N)^{-1}(c_j - \pi j)^3. \tag{A9}
\]

However, in a 3D dynamical model, \(F_{bj}\) may increase with height even above the breaking level \(z_{bj}\). In this case it is appropriate to keep \(F_j\) constant with height and to neglect the individual wave contributions to the diffusion coefficient (e.g., McFarlane 1987). We furthermore assume an exponential decay of \(D_j\) below the breaking levels as in Holton (1982). Then, letting \(\Delta z\) be the height increment between adjacent half-model layers, we have

\[
F_j(z) = \begin{cases} 
0 & \text{for } z < z_0 \\
F_{z0} & \text{for } z_0 \leq z \leq z_{bj} \\
F_{bj} & \text{for } z_{bj} < z < z_{cij} \text{ and } F_{bj} \leq F_j(z - \Delta z) \\
F_{j[z - \Delta z]} & \text{for } z_{bj} < z < z_{cij} \text{ and } F_{bj} > F_j(z - \Delta z) \\
0 & \text{for } z \geq z_{cij}.
\end{cases}
\]

\[
D_j(z) = \begin{cases} 
0 & \text{for } z \geq z_{cij} \\
F_{bj} \times \{(c - \pi j)/H + 3\partial \pi_j)/(\rho N^2) & \text{for } z_{bj} \leq z \leq z_{cij} \text{ and } F_{bj} \leq F_j(z - \Delta z) \\
F_{bj} \exp\{(z - z_{bj})/H\} & \text{for } z \leq z \leq z_{bj} \text{ and } F_{bj} > F_j(z - \Delta z) \\
0 & \text{for } z_0 \leq z < z_{bj} \\
0 & \text{for } z < z_0.
\end{cases}
\]

Equations (A10) and (A11) differ from the implementation used in BS3 in several respects. First, in BS3, \(F_j\) was set zero everywhere if no breaking level was obtained between \(z_0\) and \(z_{cij}\). Second, in BS3 we employed \(F_j[z] = F_{bj}[z_{bj}]\) for \(z_0 \leq z \leq z_{bj}\) and \(z_{bj} < z_{cij}\). However, \(F_{bj}[z_{bj}] = F_{bj}\) is valid by means of (A5), (A8), and (A9) in the continuous case only. In BS3, we also did not take the case \(F_{bj} > F_j[z - \Delta z]\) into account. Furthermore, in the present implementation, a new finite differencing scheme for frictional heating and the residual work \(-\rho^{-1}\mathbf{F} \cdot \mathbf{\partial} \mathbf{v}\) [see Eq. (11) in BS3] is employed.\(^{A1}\) This method will be described elsewhere. All these changes affect the model climatology at most slightly. However, in order to account for the reversal of the upper-level zonal-mean zonal wind in both hemispheres, we utilize a new parameter setting compared to BS3. Figure A1 summarizes our actual choices for \(k_j, c_j, c_j,\) and \(\bar{u}_j\). The total number of individual waves is \(j_0 = 14\). These waves propagate in eight equidistant azimuths. In each of the azimuths 0°, 45°, 135°, 180°, 225°, and 315°, there are two waves that differ with respect to their phase speeds \(c_j\) while having the same wavenumbers \(k_j\) and amplitudes \(\bar{u}_j\). Each of the azimuths 90° and 270° contains one wave.

\(^{A1}\)To ensure energy conservation in a numerical model, the energy deposition is specified by computing \(-\rho^{-1}\mathbf{F} \cdot \mathbf{\partial} \mathbf{v}\) and \(-\rho^{-1}\mathbf{\partial} \mathbf{F} \cdot \mathbf{\partial} \mathbf{v}\) separately.

**REFERENCES**


