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1 **Non-orographic gravity waves: representation in**
2 **climate models and effects on infrasound**

3 **D. Cugnet¹, A. de la Camara², F. Lott¹, C. Millet^{3,4}, and B. Ribstein^{4,3}**

4 ¹LMD, PSL Research University, Ecole Normale Supérieure, Paris, France

5 ²Dpto. Física de la Tierra y Astrofísica, Univ. Complutense de Madrid, Spain

6 ³CEA, DAM, DIF, F-91297 Arpaçon, France

7 ⁴CMLA, ENS Cachan, CNRS, Université Paris-Saclay, 94235 Cachan, France

Corresponding author: François Lott, flott@lmd.ens.fr

Abstract

Long-range infrasound propagation is controlled by atmospheric waveguides that extend up to the mesosphere and lower thermosphere and whose efficiency is affected by gravity waves (GWs). These GWs are not explicitly represented in the global models often used to calculate infrasound propagation because their spatial scales are well below the models resolution. These unresolved GWs also transport momentum and control in good part the large-scale circulation in the middle atmosphere. These two issues make that the GWs need to be parameterized to improve the datasets used to calculate infrasound propagation as well as in the Atmospheric General Circulation Models (AGCMs) that are used to make weather forecasts and climate predictions. These two issues gain in being treated in conjunction. From this, improved infrasound calculations could be made by using a realistic amount of GWs. In return, using infrasound records could help specifying important characteristics of the GWs that are parameterized in the climate models.

The paper presents a research framework developed to address these issues. It first presents a non-orographic GWs parameterization used and tested in a well-established AGCM, emphasizing the most recent developments, like the introduction of stochastic techniques and a better specification of the GWs sources. The significance of GWs on the global climate is then illustrated by making sensitivity tests where the frontal and convective GWs parameters are moderately changed. These changes impact the structure of the jets in the midlatitude stratosphere and the intensity of the sudden stratospheric warmings.

The paper also presents a method to calculate long-range infrasound propagation, and to incorporate the contribution of the GWs that are parameterized in the AGCM. We then show that the changes in GW parameters tested in the model also impact infrasound propagation. This makes infrasound detection a potential tool to tune GWs parameterization in large-scale models.

1 Introduction

The parameterization of gravity waves in climate models is critical for the proper representation of the circulations of both the troposphere and the middle atmosphere in Atmospheric General Circulation models (AGCMs). The orographic gravity waves sig-

nificance is described in the previous issue of this serie [Lott and Millet, 2009] whereas it is well established that the non-orographic GWs are responsible for the reversal of the meridional Temperature gradient at the mesopause [Holton, 1982] and that they are a substantially driving the quasi-biennial oscillation (QBO) in the equatorial lower stratosphere [Lindzen and Tsay, 1975].

In order to parameterize the non-orographic GWs some scheme like Alexander and Dunkerton [1999] bin the spectral domain with a large number of monochromatic waves, an approach where the GWs sources can be easily introduced [Beres et al., 2005], but that can be very expensive. Today, to circumvent this difficulty, the spectral domain is sampled by stochastic methods [Eckermann, 2011; Lott et al., 2012a; Lott and Guez, 2013; de la Camara and Lott, 2015], an approach that is further justified by the fact that (i) the mesoscale dynamics producing the GWs is chaotic and (ii) the spread of the ensemble climate predictions need to be increased by stochastic forcings [Palmer, 2012].

It is now generally recognized that long-range infrasound propagation is controlled by the stratospheric and mesospheric wind and temperature vertical profiles [Rind, 1978; Gossard and Hooke, 1975]. During the past ten years extensive simulations were carried out (Kulichkov et al. [2010]; Chunchuzov et al. [2011]; Hedlin et al. [2012]; Bertin et al. [2014], among others) confirming that these profiles yield distinct infrasound ducts, which make that a sound emitted at a given source can have several distinct arrivals. In infrasound propagation modeling, the wind and temperature profiles are usually obtained by blending results of weather prediction centers (e.g. ECMWF) with semi-empirical models of winds and Temperature for altitude around 90km and above (upper mesosphere lower thermosphere Drob et al. [2008]; Picone et al. [2002]). Nevertheless, these datasets only capture variations along quite large horizontal and vertical scales (the so-called "resolved scales"), and using these data, there are systematic missfits between predicted and observed infrasound amplitudes and waveform. One reason can be that when the acoustic ducts are borderline, in the sense that small disturbances can create or destroy a duct, the signal propagation becomes very sensitive to wind fluctuations (e.g. Bertin et al. [2014]; Dergham and Millet [2013]), which produces highly dispersed signals. In the middle atmosphere the origin of these fluctuations is likely to be due to GWs, and recent works (e.g. Lalande and Waxler [2016]; Hedlin and Drob [2014]; Drob et al. [2013]) show that their main effect is to spread the incoming infrasound signal. Without accounting for GWs, the duration of stratospheric signals is often underestimated, by a factor of 5-10. In such

72 cases also, deterministic modeling fails in predicting the waveforms and a statistical anal-
 73 ysis of sound propagation appears necessary to estimate the uncertainties. A problem
 74 with these studies is that in the middle atmosphere, the amplitude and characteristic of
 75 the GWs are still not well known. In this context we propose to test GWs field which
 76 are consistent with those reducing systematic biases in AGCMs. To specify large-scale
 77 winds that are consistent with in-situ infrasound observations, we will nevertheless fol-
 78 low *de la Camara et al.* [2014] and run the scheme off-line using the large scale winds
 79 provided by the ECMWF ERAI analysis up to the mesosphere and by an empirical model
 80 above.

81 The purpose of the present paper is to present the research framework we have re-
 82 cently built to improve infrasound calculations and gravity wave parameterizations. To
 83 that end, section 2 presents the non-orographic gravity wave scheme recently introduced
 84 in the stratospheric version of the LMDz climate model and its impact. To illustrate the
 85 significance of the parameterized waves we will make changes in the parameterization
 86 set-up and discuss their impacts. In section 3, we briefly expose one of the technique we
 87 use to evaluate infrasound propagation from a given source to a given receiver, and take
 88 the example of the Hukkakero ammunition destruction done August 18, 2016 at 12:30pm.
 89 We then expose how gravity waves fields can be included and show results for the two
 90 setups of the GWs parameterization we use.

91 2 Formalism

92 The stochastic method introduced in [Eckermann, 2011; Lott et al., 2012a] to rep-
 93 resent GWs consists in representing the subgrid-scale GWs field by a stochastic Fourier
 94 series

$$w' = \sum_{n=1}^{n=\infty} C_n \hat{w}_n(z) e^{z/2H} e^{i(\vec{k} \cdot \vec{x} - \omega t)} \quad (1)$$

95 where w' is the vertical velocity, z the vertical log-pressure coordinate, $H = 7\text{km}$ a ver-
 96 tical scale characteristic of the middle atmosphere temperatures, $\hat{w}_n(z)$ the complex ver-
 97 tical structure of a given harmonic, \vec{k} its horizontal wavenumber, ω its absolute frequency,
 98 \vec{x} and t being horizontal position and time respectively. In (1) the parameter C_n mea-
 99 sures the amplitude of a given harmonic to represent the total wave field (for instance

100 at a given place inside the model gridbox), it is often referred to as an intermittency co-
 101 efficient, its statistical interpretation makes that it satisfies

$$\sum_1^{\infty} C_n^2 = 1. \quad (2)$$

102 On top of this statistical decomposition, \vec{k} and ω are also chosen randomly within bounds
 103 covering the space and time scales we believe the model does not solve well. To describe
 104 vertical propagation, we derive from (1) an EP-flux due to the waves

$$\vec{F}^z = \rho_0(z) \overline{u'w'} = \sum_{n=1}^{n=\infty} C_n^2 \vec{F}_n^z \text{ where } \vec{F}_n^z = \rho_r \Re \left\{ \frac{\vec{u}_n \hat{w}_n^*}{2} \right\}, \quad (3)$$

105 where $\vec{F}_n^z(z)$ is the EP-flux carried by each waves, $\rho_0(z) = \rho_r e^{-z/H}$ being a character-
 106 istic density vertical profile ρ_r being constant. We then make a WKB evaluation of each
 107 harmonics introducing a diffusivity $\nu(z) = \mu/\rho_0(z)$, and limiting each wave amplitude
 108 to its statically marginal stability limit. From such consideration and from the Eliassen-
 109 Palm theorem, telling that the EP flux is constant for steady linear wave in the absence
 110 of dissipation, we construct the vertical profile of \vec{F}_n^z following the iterative rule from
 111 one model level (z) to the next above ($z + \delta z$):

$$\vec{F}_n^z(z + \delta z) = \frac{\vec{k}\Omega}{|\vec{k}||\Omega|} \Theta(\Omega(z + \delta z)\Omega(z)) \text{ Min} \left\{ |\vec{F}_n^z(z)| e^{-2\frac{\mu N^3 |\vec{k}|^3}{\rho_0 \Omega^4} \delta z}, \rho_r S_c^2 \frac{|\Omega|^3 k^{*2}}{N |\vec{k}|^4} \right\} \quad (4)$$

112 where the first fraction guarantees that the EP-flux is in the direction of the phase speed,
 113 the second term with Heaviside function $\Theta(z)$ handles critical levels, e.g. places where
 114 the intrinsic frequency $\Omega = \omega - \vec{k}\vec{U}$ changes sign (\vec{U} being the background wind), the
 115 first term in the parenthesis expresses the decay of the EP flux due to diffusion, and the
 116 last term is the saturated EP-flux. Still in (4) N is the buoyancy frequency whereas k^*
 117 is the minimum GW wavelength parameterized and S_c is a tunable parameter that con-
 118 trols the wavebreaking.

119 To relate the gravity waves to their non-orographic sources, e.g. convections and
 120 fronts, we then follow *Lott and Guez* [2013] and *de la Camara and Lott* [2015] and rep-
 121 resent the subgrid scale precipitation and vorticity by stochastic series,

$$P' = \sum_{n=1}^{\infty} C_n P_n e^{i(\vec{k} \cdot \vec{x} - \omega t)}, \quad q' = \sum_{n=1}^{\infty} C_n q_n e^{i(\vec{k} \cdot \vec{x} - \omega t)}. \quad (5)$$

122 In *Lott and Guez* [2013], the precipitation is translated into a heating over a character-
 123 istic depth Δz and integrated into a forced Taylor-Goldstein equation solved via a Green
 124 function method. It yields a "launched" momentum flux,

$$|\overline{F}^z| = G_0 \rho_r \left(\frac{RL_c}{\rho_r H c_p} \right)^2 \frac{|k|^2 e^{-\frac{N^2 |\vec{k}|^2 \Delta z_1^2}{\Omega^2}}}{N |\Omega|^3} P^2, \quad (6)$$

125 where we have assumed that the amplitude of the stochastic projection of subgrid scale
 126 precipitation onto the n^{th} harmonics equal the gridscale precipitation P : $|\hat{P}_n| = P$. In
 127 (6) the n-indices are dropped for conciseness, R is the dry air gas constant, L_c the la-
 128 tent heat of condensation, Δz_1 scales the depth of convection, and G_0 is a tunable con-
 129 stant of order 1.

130 In *de la Camara and Lott* [2015] the frontal waves are parameterized by realizing
 131 that within fronts, relative vorticity anomalies q' are always substantial, and by using
 132 closed formula for spontaneous GWs emission from PV anomalies derived in [*Lott et al.*,
 133 2010, 2012b], e.g. by writing,

$$|\overline{F}^z| = G_1 \frac{\Delta z_2}{4f} \int_0^{\infty} \rho(z') N(z') q'^2 e^{-\pi \frac{N}{U_z}} dz', \quad (7)$$

134 where Δz_2 is a tuneable depth for the PV anomalies, it is near the gridscale depth δz ,
 135 f is the Coriolis parameter, U_z is the wind shear and G_1 is a tuneable parameter, again
 136 of order 1.

137 3 Test in LMDz

138 3.1 Experimental setup

139 The LMDz version used here is one of the configurations designed for the prepara-
 140 tion of the sixth coupled model intercomparison project (CMIP6), an evolution of its
 141 CMIP5 version presented in *Hourdin et al.* [2013]. For the middle atmosphere, impor-
 142 tant differences nevertheless needs to be detailed. First the model now has 80 levels, with
 143 a top at 1Pa, and a resolution in the lower stratosphere of $\delta z \approx 1\text{km}$ near $z = 20\text{km}$,
 144 increasing smoothly to reach $\delta z \approx 1.4\text{km}$ near $z = 40\text{km}$, then the resolution increases
 145 more rapidly it equals $\delta z \approx 6\text{km}$ at the top and there are only 13 levels in the meso-

146 sphere. ($50 < z < 80\text{km}$). Second, the model has ozone specifications that are derived
 147 from observational data, rather than from the LMDz-Reprobus model as in CMIP5. The
 148 ozone data used being defined only below 1hPa, we make the O_3 mixing ratio above de-
 149 creasing rapidly as a function of pressure by using the relation,

$$\chi_{O_3}(P < 1\text{hPa}) = \chi_{O_3}(1\text{hPa}) \sqrt{\frac{P}{P(1\text{hPa})}}, \quad (8)$$

150 which reproduces qualitatively the decay with altitude of the daily Ozone in the meso-
 151 sphere used in *Hourdin et al.* [2013].

152 Concerning the parameterization of the non-orographic gravity waves, we choose
 153 the wavenumber of the waves randomly using a uniform distribution between $10^{-3} >$
 154 $k > k^* = 2 \cdot 10^{-5}$. Their intrinsic phase speed is also chosen randomly using a Gaus-
 155 sian distribution with standard deviation $U^* = 50\text{m/s}$ for the convective waves and $U^* =$
 156 30m/s for the frontal waves. Concerning the non-dimensional parameters we take $G_0 =$
 157 0.23 for the convective waves and $G_1 = 4$ for the frontal waves. Also, the launching al-
 158 titudes is around $z_l = 500\text{m}$ for the frontal waves and $z_l = 5\text{km}$ for the convective ones.
 159 Finally, we take for the saturation parameter $S_c = 0.6$ and for the diffusion $\nu = 1$. For
 160 completeness the subgrid scale orography parameters values are quite different from those
 161 given in *Lott and Millet* [2009], see their Eqs 23.1 – 23 – 6: we now take $H_{NC} = 1.$,
 162 $C_d = 0.5$, $G = 0.1$, $\beta = 0.1$, $Ri_c = 1$, and $C_l = 0.5$. Here H_{NC} represents the critical
 163 value of the non-dimensional mountain height above which the low level flow is blocked,
 164 C_d the drag coefficient applied to the blocked flow, G the gravity wave drag coefficient,
 165 β the fraction of the gravity wave drag that propagates freely in and aloft the free tro-
 166 posphere, Ri_c the critical Richardson number controlling orographic gravity waves break-
 167 ing, and C_l the mountain lift coefficient introduced in *Lott* [1999].

168 The results presented in this paper are from a 15-yrs experiment where sea surface
 169 temperature, sea-ice and ozone are imposed from climatologies constructed with 1979-
 170 2008 data. To illustrate how the GWs impact the simulated climate in the middle at-
 171 mosphere we also show an experiment with slower frontal waves, $U^* = 10\text{m/s}$. In this
 172 configuration the drags stay reasonable in the middle atmosphere if we take a slightly
 173 smaller $G_1 = 2$ and a much larger $S_c = 6.7$. In this setup the frontal wave drag is slightly
 174 smaller than in the first experiment. Conversely, we also increase the convective wave drag
 175 constant to $G_0 = 0.6$ in order to increase the convective wave drag.

3.2 Midlatitude Climatologies and sudden stratospheric warmings

The zonal mean zonal winds for January, April, August and October are shown in the 4 panels in Fig. 1a. It shows that the stratospheric winds in the winter midlatitudes are oriented eastward and exhibit well defined jets with maxima centered near 60°N in the NH and 60°S in the SH. These are the so called polar night-jets and their maximum value is about right in the NH midlatitude. Still during solstices, the winds are essentially westward in the summer hemisphere, with well define maxima in the subtropical regions above the stratopause. In the mesosphere the winds decay in amplitude as a function of altitude, also consistent with observations, and we know that these decays are only realistic if we have the right amount of gravity wave drag (see *Lott and Millet* [2009]). During equinoxes, the polar jets are eastward in both hemispheres, with amplitudes substantially smaller than the polar night jet during the solstices. In autumn, these east-erlies indicate the building up of the winter polar night characterizing the polar vortices, whereas in spring they are associated to their breakdown. The fact that the GW drag control the amplitude of these jets is clearly illustrated in Fig. 1b, which shows that the maximum jets amplitude are everywhere larger when the launched frontal GW drags are reduced.

To characterize the variability of the polar night jets, the Fig. 2a presents the evolution of the 30hPa zonal mean zonal wind at 60°South and North of the 10hPa zonal mean polar temperature at 10hPa. This diagnostics usually characterizes the onset of the so-called sudden stratospheric warmings (SSWs), corresponding to years with polar vortex breakdown, e.g. a time when the polar jet ceases and even reverses (Fig. 2a and when the winter Temperature can be 90K above its climatological value, e.g. larger than its summer mean. Again the model winter warmings are quite realistic in the NH, where they are much more pronounced than in the SH. This is of course consistent with observations: SSWs are first related to the breaking of planetary scale Rossby waves, these are much smaller in the SH because there the land-sea contrasts are much less pronounced than in the NH.

The gravity waves also affect the SSW, to a certain extent, and this is partly because they affect the zonal mean flow which controls the index of refraction of the Rossby waves. To illustrate this point Fig. 2b presents the diagnostics of SSWs with reduced frontal GWs. In them we see that the maximum winds attained in winter are almost 10m/s larger

208 than with larger frontal GW drag. The minimum polar temperature are also typically
 209 10K smaller. The amplitude of the variabilities is also increased accordingly by almost
 210 the same amount (i.e. 10K). As a "rule of thumb", we could say that the winter vari-
 211 ability, from minimum to maximum goes from the average winter minimum to the av-
 212 eraged summer maximum. As GW drag increases the former, it increases within the same
 213 amount the variability.

214 3.3 Tropics and the quasi-biennial oscillation

215 In the absence of GWs drag the LMDz model, like any other model, fails in pro-
 216 ducing a QBO. This is related to the observationnal fact that the large scale equatorial
 217 waves alone are not large enough to force the QBO: it is generally admitted that they
 218 contribute one half only, the second being the GWs. Therefore, we have tuned the con-
 219 vective GWs scheme in order that the model produces a QBO, and the results are in Fig. 3a.
 220 In it we see that the Equatorial winds at the equator alternate positive and negative phase,
 221 within a cycle that has a irregular period of almost 26 months. This is a little faster than
 222 the observed QBO (e.g., 27-28 months) but the difference can easily be corrected by slightly
 223 decreasing the GWs amplitude. Apart from this difference, the model QBO presents the
 224 characteristic features of this oscillation, like a decending phase, a relation with the an-
 225 nual cycle at its top near $1hPa$. To illustrate again that the oscillation is very sensitive
 226 to the GW drag, Fig. 3b shows the zonal mean wind at the equator when the launched
 227 convective GWs stress is increased. In this case, the periods of the oscillation falls to 16mon-
 228 thes, and the amplitude of the oscillation in the low stratosphere is also substantially in-
 229 creased.

230 4 Impact of GWs on infrasound propagation

231 4.1 Formalism

The classical approach in infrasound propagation modeling consists in solving the wave equation, often in cylindrical coordinates (r, θ, z) , where r and z are the distance from the source and altitude, respectively. While the nature of the source significantly influences the pressure field, a simpler and common way to account for the source is to introduce driving forces on the right-hand side of the wave equation. Hence, using the form of a Dirac delta function $\delta(z - z_0)\delta(r)s(t)$, where $s(t)$ is the source function, the

pressure fluctuation reads as,

$$p(z, r, t) = \frac{1}{2\pi} \int_{-\infty}^{+\infty} \tilde{s}(\omega) \tilde{p}(z; r, \omega) e^{-i\omega t} d\omega, \quad (9)$$

232 where \tilde{s} is the Fourier transform of the source. The solution of the homogeneous wave
 233 equation is then a solution of the inhomogeneous equation except at height z_0 where the
 234 field variables are not generally continuous.

The solution in normal modes (*Jensen et al.* [1994]) of the wave equation can be attacked as two quite distinct problems: (1) finding the eigenpairs in a horizontally homogeneous atmosphere of arbitrary temperature and wind distribution in the vertical direction and (2) synthesis of the waveform at a given range r and time t resulting from a specified source. While the general solution of such problems leads to the so-called leaky modes, open modes (*Budden* [1961]), these modes are usually neglected for infrasound application. Hence, in the normal mode approximation, one seeks the total pressure fluctuation as a sum of M modal contributions;

$$\tilde{p}(r, z; \omega) \approx \sum_{m=1}^{M(\omega)} \frac{\phi_m(0, \omega) \phi_m(z; \omega)}{\sqrt{k_m(\omega)}} e^{ik_m r}, \quad (10)$$

where ϕ_m and k_m satisfy the Helmholtz equation

$$\frac{\partial^2 \phi_m}{\partial z^2} + \left[\frac{\omega^2}{c(z)^2} - k_m^2 \right] \phi_m = 0, \quad (11)$$

where $c(z)$ is the effective sound speed. Here we use the effective sound speed approximation (*Godin* [2002]), in which the component of the horizontal wind speed in the direction of propagation is added to the thermodynamic sound speed,

$$c(z) = \sqrt{\gamma RT(z)} + \frac{\vec{k} \cdot \vec{u}(z)}{|\vec{k}|} \quad (12)$$

235 where the index m on the wavevector is dropped because only the projection on the di-
 236 rection of sound propagation matters: $c(z)$ is not a function of m .

237 At $z = 0$ we use with a Neumann boundary condition and the eigenfunction ϕ_m
 238 in (10) is suitably normalized (*Jensen et al.* [1994]). Finally, the signal is obtained by
 239 applying the inverse Fourier transform (9), where the integration is performed along a
 240 path lying along the real axis. In order to avoid singularities and spurious noncausal ef-
 241 fects, the frequency parameter ω is treated as a complex variable $\omega = \omega_r + i\omega_i$ (*Bertin*
 242 *et al.* [2014]), with a small positive imaginary part $\omega_i \ll 1$ and thus, the eigenvalues
 243 $k_m(\omega)$ are also complex.

Using the normal mode approach to compute the acoustic pressure field, the impact of GWs is evaluated as follows. Firstly, the large-scale vertical profiles of temperature and winds are extracted from ECMWF products, typically up to 70km. The profiles are matched with the empirical models HWM07 (*Drob et al.* [2008]) and MSIS-90 (Mass Spectrometer and Incoherent Radar Model) through interpolation functions. These two empirical models provide a statistical representation of the mesosphere and thermosphere. The resulting profiles allow computing both the effective sound speed $c(z)$, through the effective sound speed approximation, and the GW field. Secondly, the off-line version of the GW model is used to estimate the impact of GWs onto the vertical profiles. Since the scheme described in section (2) only predicts fluxes \vec{F}_n^z for randomly chosen wavevectors \vec{k}_n and frequencies ω_n , it is necessary to rederive the GW fields by applying local polarization relations and a WKB formalism, yielding,

$$(\vec{u}', T') = \Re \left\{ \sum_{n=1}^N C_n (\vec{u}_n, T_n) e^{z/2H} e^{i(\vec{k}_n \cdot \vec{x}) - \omega_n t} \right\} \quad (13)$$

where

$$(\vec{u}_n, T_n) = \Re \left\{ \left(\frac{N}{\Omega_n} \frac{\vec{k}_n}{|\vec{k}_n|}, -i \frac{N^2 H}{R \Omega_n} \right) \hat{w}_n \right\}, \quad (14)$$

and where the vertical profiles of vertical velocity,

$$\hat{w}_n = \sqrt{\frac{2 \|\vec{F}_n^z \Omega_n\|}{\rho_r N}} \exp \left(-i \int_0^z \frac{N \|\vec{k}_n\|}{\Omega_n} dz' + i \chi_n \right), \quad (15)$$

244 where χ_n is a phase that can be chosen randomly. Then, to evaluate wave fields that are
 245 consistent with the parameterization presented in section (2), we proceed stochastically
 246 and first compute $n = 100$ realizations of vertical profiles choosing \vec{k}_n and ω_n randomly.
 247 From this large number of realizations we reconstruct an ensemble of GW fields pick-
 248 ing the intermittent factors C_n randomly as well, but conserving the normalization con-
 249 dition (2).

250 4.2 A case study

We generally refer to the low frequency band of the acoustic spectrum ($2\pi N < \omega < 2\pi$) as infrasound. Man-made sources in this band are limited to large explosions and to sonic booms generated by rockets and aircrafts. Extensive investigations of such sources have been carried out by numerous authors and examples of infrasonic pressure signatures can be found in the literature. For illustration purpose we consider sound prop-

agation of regional distances, and model the source function in (9) by

$$s(t) = \cos(2\pi f_c t) \frac{1 - \cos(\pi f_c t)}{2}, \quad (16)$$

251 where $0 < t < 1/f_c$, and $s(t) = 0$ otherwise.

252 Fig. 4 shows the vertical profiles of effective sound speed as obtained by adding GWs
 253 onto the atmospheric specification provided by ECMWF (analysis, 137 levels). Due to
 254 large-scale winds, the effective sound speed presents two relative maxima, at the surface
 255 and near the stratopause (around 50km), and two relative minima, at the tropopause
 256 and the mesopause. These relative extrema define two regions, often called tropospheric
 257 and stratospheric ducts, in which $\omega^2/c^2 - k^2$ can be positive and bounded by regions
 258 where it can be negative. The efficiency of the stratospheric duct is controlled by the abil-
 259 ity of the mesopause to reflect infrasound or, in terms of normal modes, to trigger a large
 260 number of normal modes. When including GWs disturbances, an ensemble of effective
 261 celerities is produced, replacing T and \vec{u} in Eq. (12) by $T+T'$ and $\vec{u}+\vec{u}'$ respectively.
 262 Fig. 4c and Fig. 4a show the standard deviations (gray lines) of the resulting profiles pre-
 263 dicted by the non-orographic gravity wave parameterization, as obtained with the two
 264 sets of GWs parameters used in Figs. 1a and 1b respectively. As pointed out in section 2
 265 and noticed by numerous authors (e.g. *Lalande and Waxler* [2016]; *Drob et al.* [2013]),
 266 the GW field produces substantial variations in the effective sound speed all the way through
 267 the middle atmosphere. Whereas the first setup of the parameterization induces substan-
 268 tial variations at the mesopause level, the fluctuations obtained with the second setup
 269 are more important in the stratosphere and lower mesosphere.

270 Results of infrasound propagation simulations obtained with the source function
 271 defined by (16) are given in Fig. 5 in the form of space-time diagrams. The normalized
 272 Hilbert transform of signals is indicated in colors. The most obvious interpretation of
 273 signals in Fig. 5 is that each succeeding arrival is “reflected” from higher in the atmo-
 274 sphere so that its path is longer than preceding arrivals. Fig. 6a shows a typical signal,
 275 recorded at the I37NO IMS station in northern Norway on August 18, 2016 and partic-
 276 ular realizations of numerically obtained signals. In this case, the source is known to be
 277 to associated with a near-surface explosion (with a yield of approximately 20-30 t) at
 278 the Hukkakero military range in Finland (*Gibbons et al.* [2015]), located 320km away from
 279 I37NO. Such event occurs on consecutive days in August and September and provide a
 280 useful data set for the study of infrasound propagation.

281 The signal in Fig. 6a shows a function of the retarded time $t - r/c_0$ where $c_0 =$
 282 356 m.s^{-1} . The first striking result is that the sound signal arrives for positive retarded
 283 time, witnessing that it is dominated by signals that have travelled substantially in the
 284 upper atmosphere. The first arrival time is due to the tropospheric duct, and we see that
 285 it is quite small in amplitude, whereas the second one is much more substantial and due
 286 to the stratospheric duct. The next three panels below show the signals calculated via
 287 the normal mode and for a source of the form given by (16) with $f_c = 0.35\text{Hz}$. The sec-
 288 ond panel from the top gives the infrasound prediction without GWs. It shows that the
 289 model predicts well the time of arrival and the duration of the tropospheric response but
 290 largely underestimates the amplitude of the stratospheric arrival. In the presence of GWs,
 291 the separation between the tropospheric and stratospheric ducts can be much more per-
 292 meable, allowing a fraction of the signal to travel easier through the stratospheric duct.
 293 This is clearly shown in the next two panels where the green curve shows that in the pres-
 294 ence of GWs field a stronger second arrival is predicted. Interestingly, it is when the GWs
 295 phase speed are smaller in the bottom panel that the stratospheric arrivals duration is
 296 larger. It is interesting to recall that in this configuration the GWs modify more the celer-
 297 ity maxima at the stratopause according to the left panel in Fig. 4 than the effective celer-
 298 ity minima at the mesopause (right panel in Fig. 4). Finally, it is important to notice
 299 that the increase of the stratospheric arrival when there are more GWs in the stratosphere
 300 than in the mesosphere is not the result of one particular GW field. The gray shaded
 301 curves show the standard deviation of the signals out of 100 GWs fields built stochas-
 302 tically, and confirm that our infrasound prediction clearly improves when the GWs are
 303 more confined to the stratosphere.

304 5 Perspective

305 While long-range infrasound propagation modeling is a useful tool in geophysics
 306 and nuclear treaty verification, the inherent unpredictability of subgrid-scale atmosphere
 307 dynamics results in a poorly constrained propagation medium. The general approach in
 308 the infrasound research community is to superimpose a single ad'hoc "frozen" distur-
 309 bance onto the atmospheric specification, obtained through spectral models and to prop-
 310 agate infrasound in the resulting medium. Such an approach, however, fails in represent-
 311 ing the impact of small-scale dynamics which is intrinsically statistical and indeed, di-
 312 rect observations in the lower stratosphere show that the gravity wave field is very in-

313 intermittent, and often dominated by well defined large-amplitude and low-probability wave
314 packets.

315 In this paper we propose another approach, in which the GW field is described by
316 a stochastic field obtained with a multiwave stochastic parameterization of GWs cur-
317 rently in use in an AGCM. The propagation of infrasound can be obtained at a rather
318 low numerical cost with low-order reduced models (*Bertin et al.* [2014]). Such an approach
319 illustrates how two disciplines can benefit from a mutual interaction. Gravity wave fields
320 can be adapted from the climate models and superimposed onto atmospheric specifica-
321 tions in order to improve infrasound predictions. In turn, the infrasound calculations can
322 be used together with observations to constrain the GW parameterizations. In this pa-
323 per we only give preliminary results and show that a state-of-the-art GW scheme can
324 yield some improvement in infrasound prediction, provided that the phase speed of GWs
325 is much smaller than that currently in use in AGCMs. As pointed out in this paper the
326 duration of signals remains smaller than that observed. On the other hand, To improve
327 GWs parameterization, one should understand which physical processes could place more
328 low phase speed GWs in the stratosphere To improve infrasound models, one direction
329 could be to consider coupling between modes, especially when GWs fields vary in the
330 horizontal and vertical directions.

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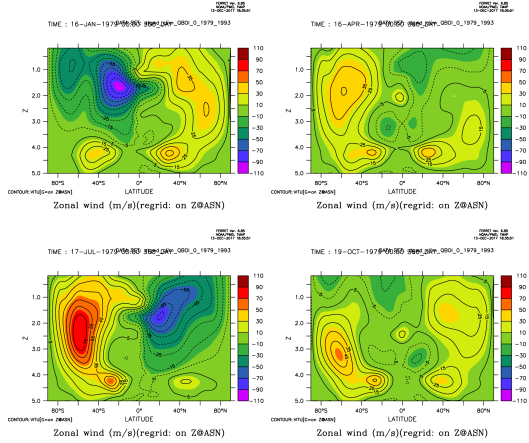
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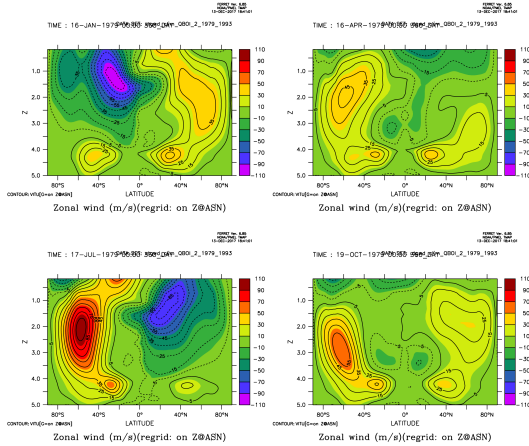
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a) Control values of the GWs drag parameter.



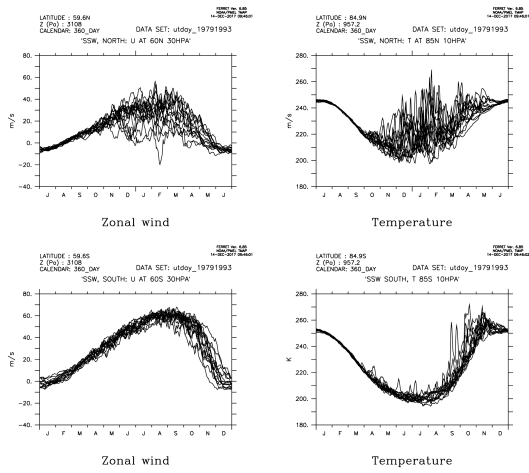
b) Modified values of the GWs drag parameter.



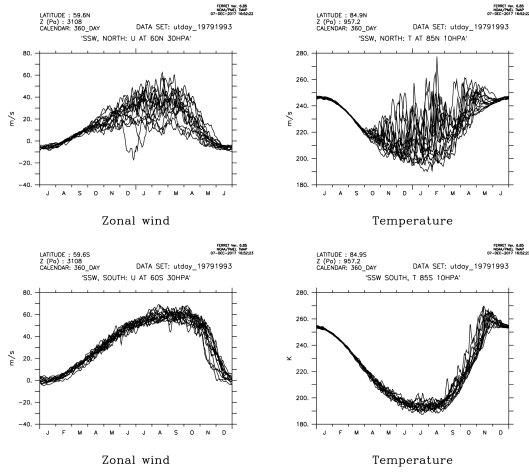
418 **Figure 1.** Climatology of the zonal wind in the control and experiments with changed GWs

419 drag parameter.

a) Control values of the GWs drag parameter.

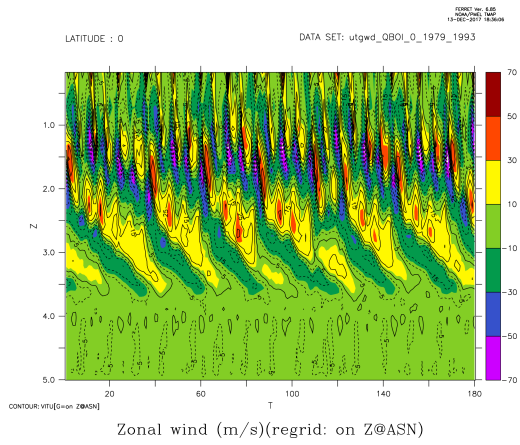


b) Modified values of the GWs drag parameter.



420 **Figure 2.** Zonal mean zonal wind at 30hPa (upper left: 60N, upper right: 60S), and polar
 421 temperature at 10hPa (upper right: 85N, lower right 85S).

a) Control values of the GWs drag parameter.



b) Modified values of the GWs drag parameter.

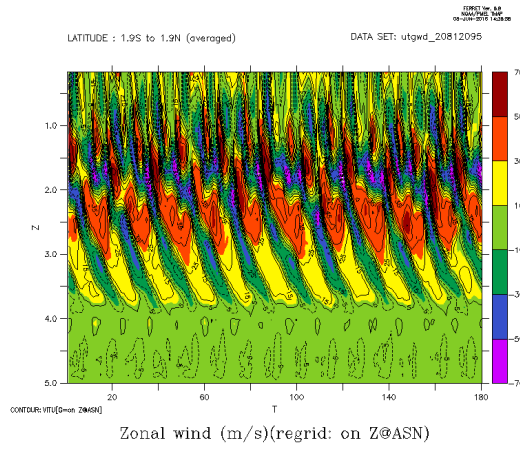
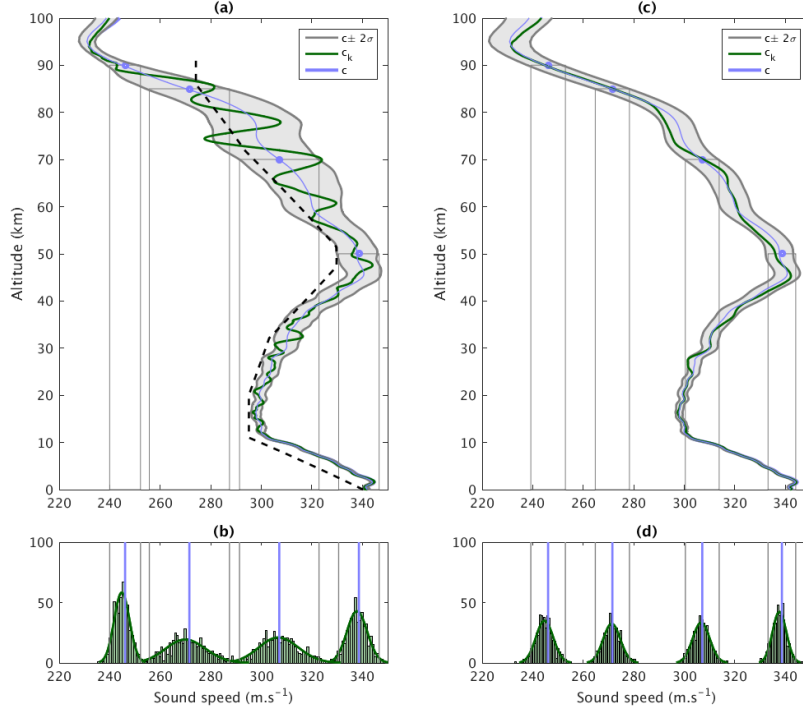
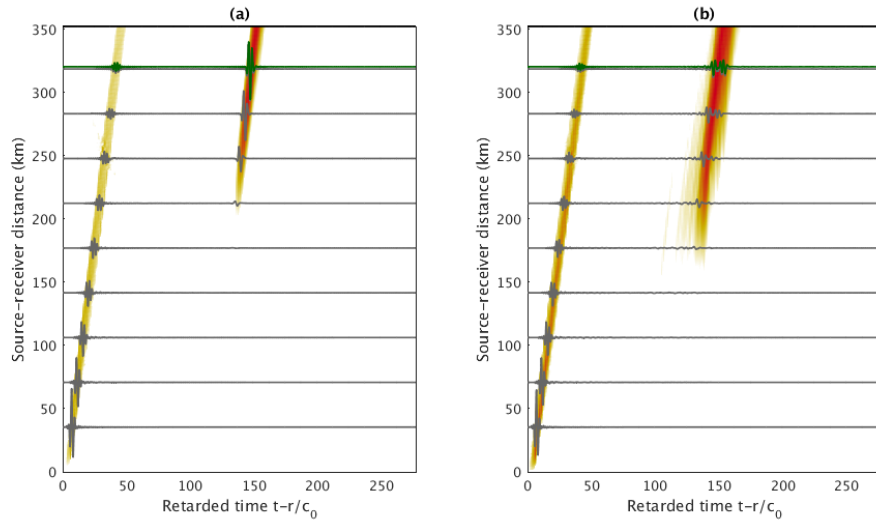


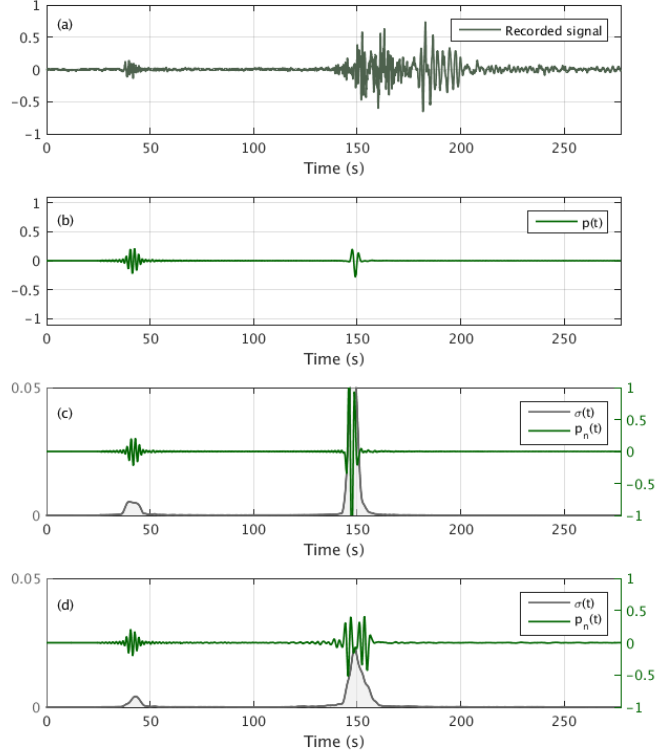
Figure 3. Zonal mean zonal wind at Equator.



423 **Figure 4.** (a,c) Profiles of effective sound speed at I37NO on 16 August, 2016 (12:30pm) and
 424 associated probability density functions (b,d) for fixed altitudes of 50, 70, 85 and 90km. The
 425 dashed black line gives the profile as obtained from the US76 empirical model. The envelope
 426 delimited by gray lines represents the standard deviation $c \pm 2\sigma$ obtained for an ensemble of 500
 427 profiles evaluated stochastically using the setup of Fig. 1 (c,d) and from a parameterization with
 428 reduced phase speed waves (a,b). The original reference profile is indicated by $c(z)$ (which is also
 429 the mean profile) and a particular realization $c_k(z)$ is given by the green line.



430 **Figure 5.** Space-time diagrams showing infrasound signals (green lines) propagating through
 431 a particular realization of the perturbed atmospheric specification. The standard deviation of the
 432 normalized Hilbert transform of signals is given by colors for an ensemble of 100 vertical profiles.
 433 (a) Results obtained using the setup of Fig. 1; (b) results obtained from a parameterization with
 434 reduced phase speed waves. The source frequency is $f_c = 0.35$ Hz.



435 **Figure 6.** Recorded signal at I37NO (a) and numerically obtained signals using the 137-levels
 436 ECMWF profile (b) and by adding gravity waves. (a) Particular signal obtained using the setup
 437 of Fig. 1 and (b) from a parameterization with reduced phase speed waves. Results are obtained
 438 for a source function defined by Eq. (16) with $f_c = 0.35$ Hz. The standard deviation σ is plot-
 439 ted in solid gray line as a function of time and normalized by the maximum amplitude of each
 440 realization.