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

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8 Abstract

Long-range infrasound propagation is controlled by atmospheric waveguides that 9 extend up to the mesosphere and lower thermosphere and whose efficiency is affected by 10 gravity waves (GWs). These GWs are not explicitly represented in the global models 11 often used to calculate infrasound propagation because their spatial scales are well be-12 low the models resolution. These unresolved GWs also transport momentum and con-13 trol in good part the large-scale circulation in the middle atmosphere. These two issues 14 make that the GWs need to be parameterized to improve the datasets used to calculate 15 infrasound propagation as well as in the Atmospheric General Circulation Models (AGCMs) 16 that are used to make weather forecasts and climate predictions. These two issues gain 17 in being treated in conjunction. From this, improved infrasound calculations could be 18 made by using a realistic amount of GWs. In return, using infrasound records could help 19 specifying important characteristics of the GWs that are parameterized in the climate 20 models. 21

The paper presents a research framework developed to address these issues. It first 22 presents a non-orographic GWs parameterization used and tested in a well-established 23 AGCM, emphasizing the most recent developments, like the introduction of stochastic 24 techniques and a better specification of the GWs sources. The significance of GWs on 25 the global climate is then illustrated by making sensitivity tests where the frontal and 26 convective GWs parameters are moderatly changed. These changes impact the struc-27 ture of the jets in the midlatitude stratosphere and the intensity of the sudden strato-28 spheric warmings. 29

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.

35 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 significance 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-biennal oscillation (QBO) in the equatorial lower stratosphere [Lindzen and Tsay, 1975].

In order to parameterize the non-orographic GWs some scheme like Alexander and 44 Dunkerton [1999] bin the spectral domain with a large number of monochromatic waves, 45 an approach where the GWs sources can be easily introduced [Beres et al., 2005], but 46 that can be very expensive. Today, to circumvent this difficulty, the spectral domain is 47 sampled by stochastic methods [Eckermann, 2011; Lott et al., 2012a; Lott and Guez, 2013; 48 de la Camara and Lott, 2015], an approach that is further justified by the fact that (i) 49 the mesoscale dynamics producing the GWs is chaotic and (ii) the spread of the ensem-50 ble climate predictions need to be increased by stochastic forcings [Palmer, 2012]. 51

It is now generally recognized that long-range infrasound propagation is controlled 52 by the stratospheric and mesospheric wind and temperature vertical profiles [Rind, 1978;53 Gossard and Hooke, 1975]. During the past ten years extensive simulations were carried 54 out (Kulichkov et al. [2010]; Chunchuzov et al. [2011]; Hedlin et al. [2012]; Bertin et al. 55 [2014], among others) confirming that these profiles yield distinct infrasound ducts, which 56 make that a sound emitted at a given source can have several distinct arrivals. In infra-57 sound propagation modeling, the wind and temperature profiles are usually obtained by 58 blending results of weather prediction centers (e.g. ECMWF) with semi-empirical mod-59 els of winds and Temperature for altitude around 90km and above (upper mesosphere 60 lower thermosphere Drob et al. [2008]; Picone et al. [2002]). Nevertheless, these datasets 61 only capture variations along quite large horizontal and vertical scales (the so-called "re-62 solved scales"), and using these data, there are systematic missfits between predicted and 63 observed infrasound amplitudes and waveform. One reason can be that when the acous-64 tic ducts are borderline, in the sense that small disturbances can create or destroy a duct, 65 the signal propagation becomes very sensitive to wind fluctuations (e.g. Bertin et al. [2014]; 66 Dergham and Millet [2013]), which produces highly dispersed signals. In the middle at-67 mosphere the origin of these fluctuations is likely to be due to GWs, and recent works 68 (e.g. Lalande and Waxler [2016]; Hedlin and Drob [2014]; Drob et al. [2013]) show that 69 their main effect is to spread the incoming infrasound signal. Without accounting for GWs, 70 the duration of stratospheric signals is often underestimated, by a factor of 5-10. In such 71

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

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

91 2 Formalism

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

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

where w' is the vertical velocity, z the vertical log-pressure coordinate, H = 7km a vertical scale characteristic of the middle atmosphere temperatures, $\hat{w}_n(z)$ the complex vertical structure of a given harmonic, \vec{k} its horizontal wavenumber, ω its absolute frequency, \vec{x} and t being horizontal position and time respectively. In (1) the parameter C_n measures 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

$$\sum_{1}^{\infty} C_n^2 = 1. \tag{2}$$

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

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

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

$$\vec{F}_{n}^{z}(z+\delta z) = \frac{\vec{k}\Omega}{|\vec{k}||\Omega|} \Theta\left(\Omega(z+\delta z)\Omega(z)\right)$$
$$\operatorname{Min}\left\{|\overline{F}_{n}^{z}(z)|e^{-2\frac{\mu N^{3}|\vec{k}|^{3}}{\rho_{0}\Omega^{4}}\delta z}, \rho_{r}S_{c}^{2}\frac{|\Omega|^{3}k^{*2}}{N|\vec{k}|^{4}}\right\}$$
(4)

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

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

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

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

$$|\overline{F}^{z}| = G_{0}\rho_{r} \left(\frac{RL_{c}}{\rho_{r}Hc_{p}}\right)^{2} \frac{|k|^{2}e^{-\frac{N^{2}|\vec{k}|^{2}\Delta z_{1}^{2}}{\Omega^{2}}}}{N|\Omega|^{3}}P^{2},$$
(6)

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

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

$$|\overline{F}^{z}| = G_{1} \frac{\Delta z_{2}}{4f} \int_{0}^{\infty} \rho(z') N(z') q'^{2} e^{-\pi \frac{N}{U_{z}}} dz',$$
(7)

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

¹³⁷ 3 Test in LMDz

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3.1 Experimental setup

The LMDz version used here is one of the configurations designed for the preparation of the sixth coupled model intercomparison project (CMIP6), an evolution of its CMIP5 version presented in *Hourdin et al.* [2013]. For the middle atmosphere, important differences nevertheless needs to be detailed. First the model now has 80 levels, with a top at 1Pa, and a resolution in the lower stratosphere of $\delta z \approx 1$ km near z = 20km, increasing smoothly to reach $\delta z \approx 1.4$ km near z = 40km, then the resolution increases more rapidly it equals $\delta z \approx 6$ km at the top and there are only 13 levels in the mesosphere. (50 < z < 80km). Second, the model has ozone specifications that are derived from observational data, rather than from the LMDz-Reprobus model as in CMIP5. The ozone data used being defined only below 1hPa, we make the O_3 mixing ratio above decreasing rapidly as a function of pressure by using the relation,

$$\chi_{O_3}(P < 1hPa) = \chi_{0_3}(1hPa) \sqrt{\frac{P}{P(1hPa)}},$$
(8)

which reproduces qualitatively the decay with altitude of the daily Ozone in the mesosphere used in *Hourdin et al.* [2013].

Concerning the parameterization of the non-orographic gravity waves, we choose 152 the wavenumber of the waves randomly using a uniform distribution between $10^{-3} >$ 153 $k > k^* = 2.10^{-5}$. Their intrinsic phase speed is also chosen randomly using a Gaus-154 sian distribution with standard deviation $U^* = 50$ m/s for the convective waves and $U^* =$ 155 30 m/s for the frontal waves. Concerning the non-dimensional parameters we take $G_0 =$ 156 0.23 for the convective waves and $G_1 = 4$ for the frontal waves. Also, the launching al-157 titudes is around $z_l = 500m$ for the frontal waves and $z_l = 5km$ for the convective ones. 158 Finally, we take for the saturation parameter $S_c = 0.6$ and for the diffusion $\nu = 1$. For 159 completeness the subgrid scale orography parameters values are quite different from those 160 given in Lott and Millet [2009], see their Eqs 23.1 - 23 - 6: we now take $H_{NC} = 1$. 161 $C_d = 0.5, G = 0.1, \beta = 0.1, Ri_c = 1, \text{ and } C_l = 0.5.$ Here H_{NC} represents the critical 162 value of the non-dimensional mountain height above which the low level flow is blocked, 163 C_d the drag coefficient applied to the blocked flow, G the gravity wave drag coefficient, 164 β the fraction of the gravity wave drag that propagates freely in and aloft the free tro-165 posphere, Ri_c the critical Richardson number controlling orographic gravity waves break-166 ing, and C_l the mountain lift coefficient introduced in Lott [1999]. 167

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

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3.2 Midlatitude Climatologies and sudden stratospheric warmings

The zonal mean zonal winds for January, April, August and October are shown in 177 the 4 panels in Fig. 1a. It shows that the stratospheric winds in the winter midlatitudes 178 are oriented eastward and exhibit well defined jets with maxima centered near 60^{0} N in 179 the NH and 60° S in the SH. These are the so called polar night-jets and their maximum 180 value is about right in the NH midlatitude. Still during solstices, the winds are essen-181 tially westward in the summer hemisphere, with well define maxima in the subtropical 182 regions above the stratopause. In the mesosphere the winds decay in amplitude as a func-183 tion of altitude, also consistent with observations, and we know that these decays are 184 only realistic if we have the right amount of gravity wave drag (see Lott and Millet [2009]). 185 During equinoxes, the polar jets are eastward in both hemispheres, with amplitudes sub-186 stantially smaller than the polar night jet during the solstices. In autumn, these east-187 erlies indicate the building up of the winter polar night characterizing the polar vortices, 188 whereas in spring they are associated to their breakdown. The fact that the GW drag 189 control the amplitude of these jets is clearly illustrated in Fig. 1b, which shows that the 190 maximum jets amplitude are everywhere larger when the launched frontal GW drags are 191 reduced. 192

To characterize the variability of the polar night jets, the Fig. 2a presents the evo-193 lution of the 30hPa zonal mean zonal wind at 60°South and North of the 10hPa zonal 194 mean polar temperature at 10hPa. This diagnostics usually characterizes the onset of 195 the so-called sudden stratospheric warmings (SSWs), corresponding to years with po-196 lar vortex breakdown, e.g. a time whene the polar jet ceases and even reverses (Fig. 2a 197 and when the winter Temperaure can be 90K above its climatological value, e.g. larger 198 than its summer mean. Again the model winter warmings are quite realistic in the NH, 199 where they are much more pronounced than in the SH. This is of course consistent with 200 observations: SSWs are first related to the breaking of planetary scale Rossby waves, these 201 are much smaller in the SH because there the land-sea contrasts are much less pronounced 202 than in the NH. 203

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

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

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3.3 Tropics and the quasi-biennal oscillation

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

²³⁰ 4 Impact of GWs on infrasound propagation

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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,

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

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where \tilde{s} is the Fourier transform of the source. The solution of the homogeneous wave equation is then a solution of the inhomogeneous equation except at height z_0 where the 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;

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

where ϕ_m and k_m satisfy the Helmholtz equation

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

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,

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

where the index m on the wavevector is dropped because only the projection on the direction of sound propagation matters: c(z) is not a function of m.

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

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

where

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

and where the vertical profiles of vertical velocity,

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

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

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

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

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

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

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

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

304 5 Perspective

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

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termittent, and often dominated by well defined large-amplitude and low-probability wavepackets.

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

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Figure 1. Climatology of the zonal wind in the control and experiments with changed GWs 418 drage parameter. 419



200.

180.

MJJA

Temperature

-40.

м

Zonal wind

N

a) Control values of the GWs drag parameter.

Figure 2. Zonal mean zonal wind at 30hPa (upper left: 60N, upper right: 60S), and polar
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.



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



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



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