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1                   **Mesoscale modeling of Venus' bow-shape waves**

2                   Maxence Lefèvre<sup>1</sup>, Aymeric Spiga<sup>1,2</sup>, and Sébastien Lebonnois<sup>1</sup>

3                   <sup>1</sup>Laboratoire de Météorologie Dynamique (LMD/IPSL), Sorbonne Université,  
4                   Centre National de la Recherche Scientifique, École Polytechnique, École  
5                   Normale Supérieure, Paris, France

6                   <sup>2</sup>Institut Universitaire de France, Paris, France

7                   submitted to Icarus

## 8 Abstract

9 The *Akatsuki* instrument LIR measured an unprecedented wave feature at the top of Venusian  
10 cloud layer. Stationary bow-shape waves of thousands of kilometers large lasting several Earth  
11 days have been observed over the main equatorial mountains. Here we use for the first time a  
12 mesoscale model of the Venus's atmosphere with high-resolution topography and fully coupled  
13 interactive radiative transfer computations. Mountain waves resolved by the model form large-  
14 scale bow shape waves with an amplitude of about 1.5 K and a size up to several decades of  
15 latitude similar to the ones measured by the *Akatsuki* spacecraft. The maximum amplitude of  
16 the waves appears in the afternoon due to an increase of the near-surface stability. Propagating  
17 vertically the waves encounter two regions of low static stability, the mixed layer between  
18 approximately 18 and 30 km and the convective layer between 50 and 55 km. Some part of the  
19 wave energy can pass through these regions via wave tunneling. These two layers act as wave  
20 filter, especially the deep atmosphere layer. The encounter with these layers generates trapped  
21 lee waves propagating horizontally. No stationary waves is resolved at cloud top over the polar  
22 regions because of strong circumpolar transient waves, and a thicker deep atmosphere mixed  
23 layer that filters most of the mountain waves.

## 24 1 Intro

25 The influence of the topography on the Venusian atmosphere, especially on the dynamics of the  
26 cloud layer, is still not fully understood. The *VeGa* balloons campaign was the first measure-  
27 ment of the impact of the surface on the atmospheric dynamics and demonstrated an increase  
28 of the vertical wind above Aphrodite Terra, speculated to be linked to the propagation of oro-  
29 graphic gravity waves (Blamont et al., 1986). Numerical modeling was then used to assess the  
30 possibility for such kind of wave to emerge Young et al. (1987, 1994). These modeling efforts  
31 showed that gravity waves generated by the topography can propagate up to the cloud layer,  
32 even in the observed conditions of strong vertical variations of zonal wind and stability. Addi-  
33 tional observations of the interactions between the surface and the atmosphere have been made  
34 with the *Venus Express* mission (Bertaux et al., 2016). Correlations between the zonal wind at  
35 the top of the cloud and the underlying topography has been evidenced by UV measurements

36 with the Venus Monitoring Camera, and interpreted as the result of stationary gravity waves.  
37 A water minimum has also been measured with *Venus Express* instrument SPICAV at cloud  
38 top above Aphrodite Terra (Fedorova et al., 2016) possibly linked to an interaction between  
39 the surface and the cloud layer.

40 Using the LIR instrument on board the *Akatsuki* mission, Fukuhara et al. (2017) discov-  
41 ered large-scale stationary bow-shaped oscillations at the top of the cloud above Aphrodite  
42 Terra. This bow-shape signature, extending over  $60^\circ$  of latitude (about 10 000 km across),  
43 was observed during 5 days. Similar signatures were also reported above the main equato-  
44 rial topographic features (Kouyama et al., 2017, e.g., Atla and Beta Regio). Those signatures  
45 were interpreted as stationary orographic gravity waves. The bow-shape wave above Aphrodite  
46 Terra has been observed in the afternoon, with a maximum amplitude close to the evening  
47 terminator. The cloud-top signatures above the other Venusian mountains are visible during  
48 the afternoon, with a maximum close to the evening terminator. Cloud-top signatures associ-  
49 ated with the underlying topography were also observed above Beta Regio with the instrument  
50 *Akatsuki*/IR2 (2.02  $\mu\text{m}$  wavelength, Satoh et al., 2017) as well as with *Akatsuki* UV imager  
51 above Aphrodite Terra, Atla and Beta Regio (Kitahara et al., 2019). Stationary features have  
52 also been measured by the *Venus Express* instrument VIRTIS above the main topographical  
53 obstacles in the southern hemisphere in the nightside with no apparent dependence with the  
54 local time (Peralta et al., 2017).

55 To investigate the atmospheric dynamics of those observed phenomena, two modeling ap-  
56 proaches were adopted: one by Fukuhara et al. (2017) using a very idealized (i.e. no represen-  
57 tation of physical processes and topography) high-resolution Global Circulation Model (GCM),  
58 with an arbitrary perturbation as the source of the waves and one by Navarro et al. (2018) us-  
59 ing a subgrid-scale parameterization for gravity waves in a low-resolution GCM with complete  
60 physics (the IPSL Venus GCM described in Garate-Lopez and Lebonnois, 2018). In the study  
61 by Navarro et al. (2018), the parametric representation of the orographic gravity waves in a  
62 GCM demonstrated the link between the topography and the bow-shaped waves. In addition,  
63 the impact of the large-scale mountain waves detected by *Akatsuki* on the global-scale dynamics  
64 was shown to induce, along with the thermal tide and baroclinic waves, a significant change in  
65 the rotation rate of the solid body, with variability of the order of minutes.

66 Despite those recent modeling studies, open questions remain on the source, propagation  
67 and impact of the large-scale gravity waves evidenced by Akatsuki (Fukuhara et al., 2017;  
68 Kouyama et al., 2017). Notably, Navarro et al. (2018) used a parameterization for orographic  
69 gravity waves on Venus and did not resolve the complete emission and propagation of gravity  
70 waves from the surface to Venus' cloud top, which is not possible with current GCMs for Venus.  
71 Following the methodology described in Lefèvre et al. (2018) to study the turbulence and gravity  
72 waves in the cloud layers, we propose a method combining high-resolution atmospheric dynamics  
73 with detailed physics of the Venusian atmosphere to resolve the emission and propagation of  
74 mountain waves in a realistic atmosphere. In other words, in order to address the Venusian bow-  
75 shaped mountain waves, we developed the first Venusian mesoscale model. With high-resolution  
76 topography from *Magellan* data, we can model the atmospheric dynamics of a limited area of  
77 Venus using the dynamical core of the Weather Research and Forecast (WRF, Skamarock  
78 and Klemp, 2008). Following a method used on Mars by Spiga and Forget (2009), the WRF  
79 dynamical core is interfaced with the physics of the IPSL Venus GCM (Lebonnois et al., 2010;  
80 Garate-Lopez and Lebonnois, 2018) to be able to compute radiative transfer and subgrid-scale  
81 turbulent mixing coupled in real-time with the dynamical integrations.

82 In this paper, using our new mesoscale model for Venus, we carry out simulations in areas  
83 surrounding the main mountains of the Venusian equatorial belt at various local times, in  
84 order to understand the generation and vertical propagation of the waves and to interpret the  
85 signal detected by Akatsuki at the top of the cloud. Our new mesoscale model for Venus can  
86 potentially be used for many applications other than studying the bow-shaped mountain waves,  
87 e.g. near-surface slope winds (Lebonnois et al., 2018), polar meteorology Garate-Lopez et al.  
88 (2015), mesoscale structures in the vicinity of the super-rotating jet (Horinouchi et al., 2017).

89 This paper is organized as follows. Our mesoscale model for Venus is described in Section 2.  
90 In Section 3, the emission of resolved gravity waves is presented. The resulting top-of-the-cloud  
91 signal above Aphrodite Terra is developed in Section 4, and above Atla Regio and Beta Region  
92 in Section 5. In Section 6, mountain waves in the polar regions are discussed. Our conclusions  
93 are summarized in Section 7.

## 94 **2 The LMD Venus mesoscale model**

### 95 **2.1 Dynamical core**

96 The dynamical core of our LMD Venus mesoscale model is based on the Advanced Research  
97 Weather-Weather Research and Forecast (hereafter referred to as WRF) terrestrial model (Ska-  
98 marock and Klemp, 2008). This methodology is similar to the one adopted for the LMD Mars  
99 mesoscale model (see Spiga and Forget (2009) for a reference publication of this model and  
100 Spiga and Smith (2018) for the most up-to-date description). The WRF dynamical core in-  
101 tegrates the fully compressible non-hydrostatic Navier-Stokes equations on a defined area of  
102 the planet. The conservation of the mass, momentum, and entropy is ensured by an explic-  
103 itly conservative flux-form formulation of the fundamental equations (Skamarock and Klemp,  
104 2008), based on mass-coupled meteorological variables (winds and potential temperature). To  
105 ensure the stability of the model, and given the typical horizontal scales aimed at in our studies,  
106 the fundamental equations are integrated under the hydrostatic approximation, available as a  
107 runtime option in WRF.

### 108 **2.2 Coupling with complete physical packages for Venus**

109 The radiative heating rates, solar and IR, are calculated using the IPSL Venus GCM radiative  
110 transfer scheme (Lebonnois et al., 2015). This setting is similar to the “online” mode described  
111 in Lefèvre et al. (2018). The time step ratio between the dynamical and physical integrations  
112 is set to 250, as a trade-off between computational efficiency and the requirement that the  
113 physical timestep is significantly less than the typical radiative timescale for Venus.

114 The version of the Venus radiative transfer model used in our mesoscale model is the same  
115 as in the version of the IPSL Venus GCM described in Garate-Lopez and Lebonnois (2018).  
116 The infrared (IR) transfer uses Eymet et al. (2009) net-exchange rate (NER) formalism: the  
117 exchanges of energy between the layers are computed before the dynamical simulations are  
118 carried out by separating temperature-independent coefficients from the temperature-dependent  
119 Planck functions of the different layers. These temperature-independent coefficients are then  
120 used in the mesoscale simulations to compute the infrared cooling rates of each layer. The solar  
121 rates are based on Haus et al. (2015) computations: look-up tables of vertical profiles of the

122 solar heating rate as a function of solar zenith angle are read, before being interpolated on the  
123 vertical grid adopted for mesoscale integrations.

124 The cloud model used in this study is based on the Haus et al. (2014) and Haus et al.  
125 (2015) models derived from retrievals carried out with *Venus Express* instruments. The cloud  
126 structure is latitude-dependent, the cloud top varies from 71 km to 62 km between the Equator  
127 and the poles Haus et al. (2014). This latitudinal variation of the cloud takes the form of five  
128 distinct latitude intervals:  $0^\circ$  to  $50^\circ$ ,  $50^\circ$  to  $60^\circ$ ,  $60^\circ$  to  $70^\circ$ ,  $70^\circ$  to  $80^\circ$  and  $80^\circ$  to  $90^\circ$ . For each  
129 latitudinal intervals different NER-coefficient matrices are computed on the vertical levels of  
130 the model, ranging from the surface to roughly 100 km altitude. The cloud structure used for  
131 the calculations is fixed prior to the simulation and does not evolve with time.

132 Since the horizontal grid spacing for the mesoscale simulations is set to several tens of  
133 kilometer, the convective turbulence in the planetary boundary layer and the cloud layer is not  
134 resolved. Therefore, similarly to what is done in GCMs, our mesoscale model uses subgrid-  
135 scale parameterizations for turbulent mixing. As in the IPSL Venus GCM, for mixing by  
136 smaller-scale turbulent eddies, the formalism of Mellor and Yamada (1982) is adopted, which  
137 calculates with a prognostic equation the turbulent kinetic energy and mixing length. For  
138 mixing by larger-scale turbulent plumes (such as those resolved by Large-Eddy Simulations,  
139 see Lefèvre et al., 2017, 2018), a simple dry convective adjustment is used to compute mixed  
140 layers in situations of convectively-unstable temperature profiles. In the physics (same as in the  
141 GCM), the dependency of the heat capacity with temperature  $c_p(T)$  is taken into account, but  
142 in the dynamical core, we use a constant heat capacity as in Lefèvre et al. (2018), with a mean  
143 value of  $c_p = 1000 \text{ J kg}^{-1} \text{ K}^{-1}$  suitable for the vertical extent of our model from the surface  
144 to 100 km altitude. The initial state and the boundary conditions of the mesoscale domain,  
145 detailed in the next section, are calculated using a variable  $c_p$  to ensure a realistic forcing of  
146 the mesoscale model by the large-scale dynamics. Therefore the impact of the constant value  
147 of  $c_p$  is minimum in our configuration because the equations of the WRF dynamical core are  
148 formulated in potential temperature and the conversion from the dynamics to the physics (and  
149 vice versa) between potential temperature and temperature is done with a variable  $c_p(T)$ .

## 150 2.3 Simulation settings

151 The topography used in our mesoscale model is presented in Figure 1, it is based on *Magellan*  
152 data (Ford and Pettengill, 1992) for the majority of the surface, with *Pioneer Venus* data used  
153 to fill the blank spots wherever needed (Pettengill et al., 1980). The dataset has a resolution of  
154 8192 points along the longitude and 4096 points along the latitude. Given the detection of bow-  
155 shaped features by the *Akatsuki* spacecraft (Kouyama et al., 2017), we choose three domains  
156 of interest : Aphrodite Terra, Atla Regio and Beta Regio. These three mesoscale domains  
157 has been computed with a horizontal resolution of 40 km for Aphrodite Terra, 30 km for Beta  
158 Region and 15 km for Atla Regio. Given the horizontal resolutions involved, the dynamical  
159 timestep for mesoscale integrations is set between 8 and 10 s (typical for those horizontal  
160 resolutions, see Spiga and Forget, 2009). The horizontal domains and resolutions have been  
161 chosen to enclose the whole latitudinal extent of the bow-shaped waves, while keeping a feasible  
162 computing time. The vertical grid is composed of 150 levels from the ground to 100 km, with  
a similar distribution as the LMD Venus LES model (Lefèvre et al., 2018).

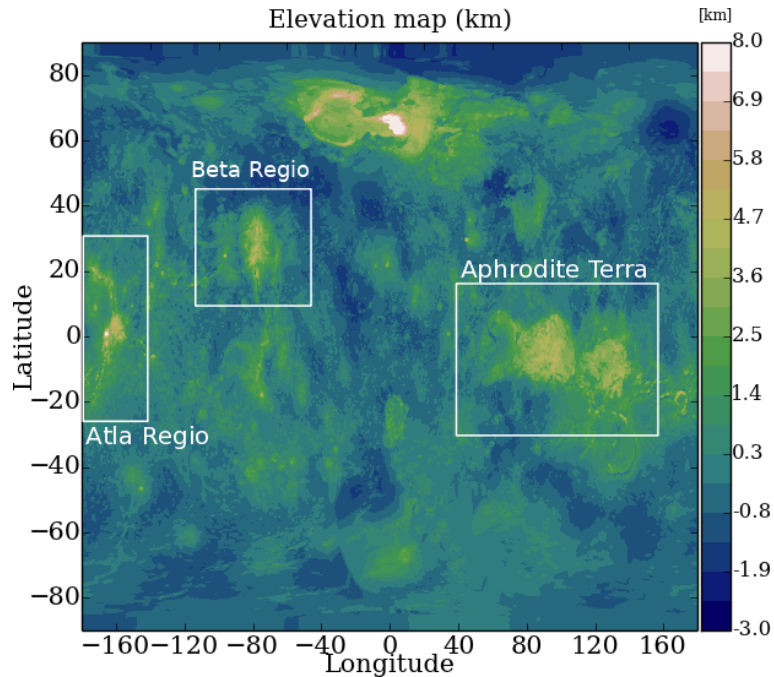


Figure 1: Topographic map of Venus used for the mesoscale simulations, from *Magellan* (Ford and Pettengill, 1992) and *Pioneer Venus* (Pettengill et al., 1980) datasets. The different domains used in the mesoscale simulations are indicated : Aphrodite Terra, Beta Regio and Atla Terra.

163

164 The horizontal boundary conditions have been chosen as ‘specified’, i.e. meteorological fields  
165 are extracted from IPSL Venus GCM simulations (detailed in Garate-Lopez and Lebonnois,

166 2018) and interpolated both on the vertical grid, accounting for the refined topography of the  
167 mesoscale, and on the temporal dimension, accounting for the evolution of those fields over the  
168 low dynamical timestep in mesoscale integrations. This ensures that the planetary-scale super-  
169 rotation of the Venusian atmosphere is well represented. The mesoscale simulations presented  
170 in this study are performed using an update frequency of a 1/100 Venus day, which enables a  
171 correct representation of the large-scale variability simulated by the GCM at the boundaries  
172 of the mesoscale domain. Between the mesoscale domain and the specified boundary fields, a  
173 relaxation zone is implemented in order to allow for the development of the mesoscale circu-  
174 lations inside the domain, while keeping prescribed GCM fields at the boundaries (Skamarock  
175 and Klemp, 2008). In this study, the number of relaxation grid points is set to 5. An expo-  
176 nential function is used to reach a smooth transition between the domain and the specified  
177 boundary fields; the coefficient is set to 1, a value used for terrestrial and Martian applications  
178 that appeared suitable for the Venusian case as well. At the top of the mesoscale model, a diffu-  
179 sive Rayleigh damping layer is applied to avoid any spurious reflection of vertically-propagating  
180 gravity waves. The damping coefficient is set to 0.01 and the depth of the damping layer to  
181 5 km.

182 The initialization of the meteorological fields for the mesoscale domain are, similarly to  
183 boundary conditions, extracted from the Garate-Lopez and Lebonnois (2018) GCM. First the  
184 fields are interpolated horizontally to the mesoscale refined-resolution grid, then interpolated  
185 vertically on the mesoscale vertical grid – accounting for high-resolution topography. To extrap-  
186 olate the high-resolution features, the same methodology is used as in the LMD Mars Mesoscale  
187 Model (Spiga and Forget, 2009).

188 Venus’s rotation is retrograde, but the WRF dynamical core has been built for Earth ap-  
189 plications, and therefore assuming implicitly prograde rotation. In order to solve this issue, as  
190 is the case for the IPSL GCM runs (Lebonnois et al., 2010), the GCM fields are turned upside  
191 down and the meridional wind is multiplied by minus 1. At the post-processing stage, the fields  
192 are turned upside down again and the meridional wind is again multiplied by minus 1, to obtain  
193 the diagnostics enclosed in this paper.

### 194 3 Generation and propagation of orographic gravity waves

195 To discuss the orographic wave generation, we focus first on Atla Regio which facilitates the  
 196 visualization of the phenomena given its very sharp mountains. Figure 2 shows the domain  
 197 chosen for Atla Regio.

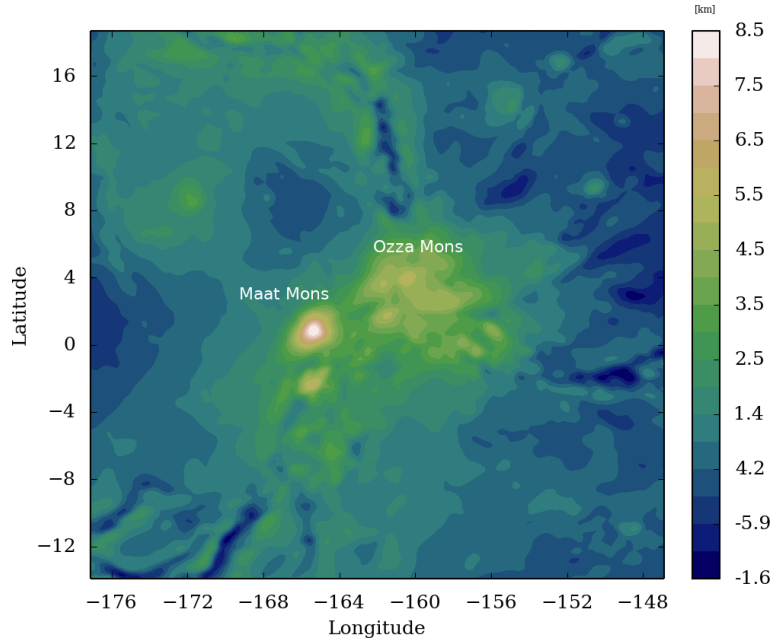


Figure 2: Elevation map (km) of the selected domain for Atla Regio with a resolution of 15 km.

198 Figure 3 shows a vertical cross-section of the vertical wind ( $\text{m s}^{-1}$ ) at  $1^\circ$  of latitude, be-  
 199 tween the surface and approximately 55 km in the beginning of afternoon. Contours represent  
 200 potential temperature. For this figure 3, and all the similar cross-sections shown in this paper,  
 201 the zonal wind comes for the right side of the plot, i.e. wind is blowing westward. The Pioneer  
 202 Venus probe measured horizontal wind ( $u$  and  $v$ ) from the ground to 3 km between 0.1 and 1.5  
 203  $\text{m s}^{-1}$  (Schubert et al., 1980). In the GCM, the horizontal wind on the same vertical extent is  
 204 between 0.1 and about 3.0  $\text{m s}^{-1}$ . The horizontal wind is of the same order of magnitude as  
 205 in-situ measurements although slightly overestimated.

206 Strong vertical winds are visible above the two main mountains, Maat Mons at  $-165^\circ$  of  
 207 longitude and Ozza Mons at  $-156^\circ$  of longitude. The interaction between the incoming zonal  
 208 flow and those sharp topographical obstacles leads to the generation of gravity waves. Near the  
 209 surface, the value of the dimensionless mountain height  $H_d = H N_B / u$  can reach value close to  
 210 1, where  $H$  is the maximum mountain height,  $N_B$  is Brunt-Väisälä frequency and  $u$  the zonal  
 211 wind, meaning that the flow is in a non-linear regime (Durrán, 2003). The waves propagate

212 vertically and encounter two regions of low static stability, the mixed layer between 18 and  
 213 35 km and the convective layer between 48 and 52 km. Above, the waves propagate into the  
 214 stratified layers. The vertical wavelength of the waves is around 30 km.

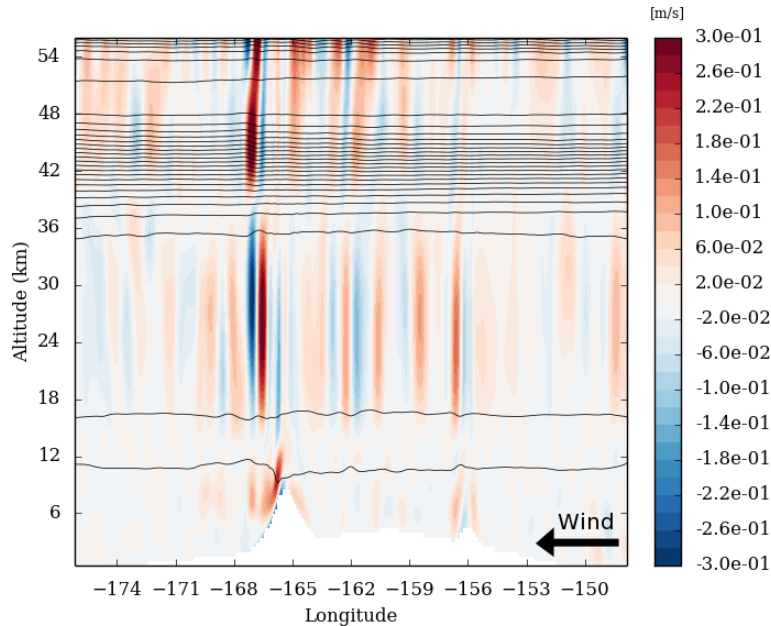


Figure 3: Vertical cross-section of the vertical wind ( $\text{m s}^{-1}$ ) at  $1^\circ$  of latitude, between the surface and approximately 55 km in the beginning of afternoon. Contours represent the potential temperature. Direction of the zonal is indicated with the black arrow.

215 The impact of these two near-neutral low-stability regions on the vertical propagation of  
 216 the wave can be understood via the Scorer parameter ( $\text{km}^{-1}$ ) (Scorer, 1949) :  $l^2 = \frac{N_B^2}{u^2} - \frac{1}{u} \frac{d^2 u}{dz^2}$ .  
 217 This parameter represents the minimum vertical wavelength of propagation. Figure 4 shows  
 218 the vertical cross-section of the Scorer parameter, and its domain-averaged vertical profile, at  
 219 the same location as Figure. 3.

220 Beyond 12 km altitude, the Scorer parameter decreases strongly up to 18 km and has very  
 221 small values, sometimes negative, from 18 to 35 km. Then it increases up to 42 km, and de-  
 222 creases until reaching the convective layer at 48 km. Inside this convective layer, the Scorer  
 223 parameter is again very small. Above the convective layer, the Scorer parameter increases  
 224 through the stable atmosphere. These two regions of low Scorer parameter indicate the pre-  
 225 dominance of trapped lee waves that propagate horizontally. The horizontally-propagating  
 226 trapped waves are visible in Figure 3 within layers exhibiting low Scorer parameter, between  
 227 18 and 35 km and between 48 and around 54 km. An additional vertical propagation of those  
 228 trapped waves is visible at some longitudes. This phenomenon is similar to the leakage of  
 229 trapped lee waves into the stratosphere on Earth (Brown, 1983). On Venus, the horizontal

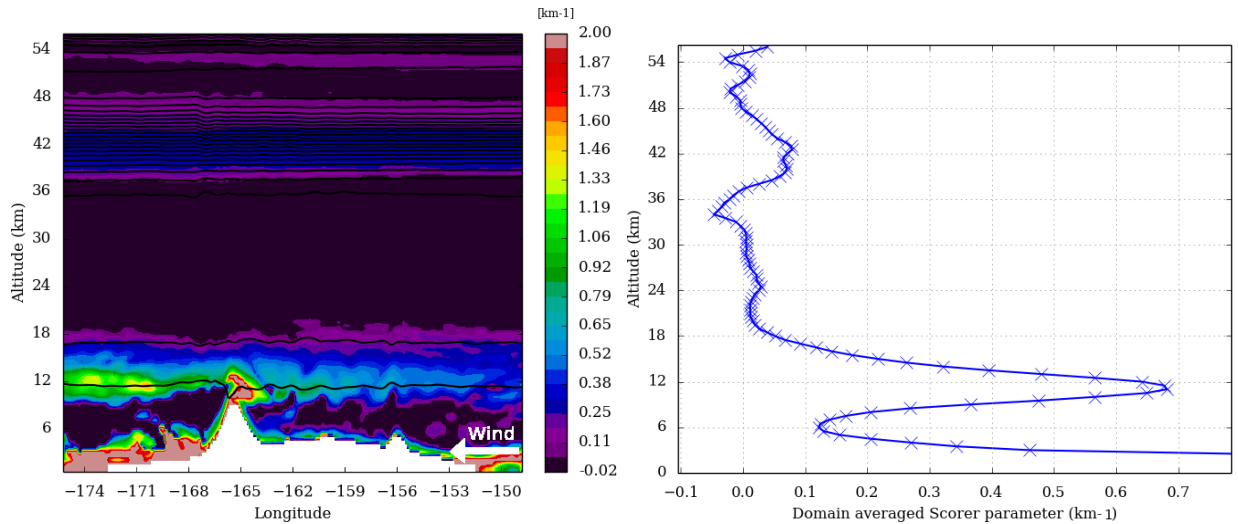


Figure 4: Left : Instantaneous vertical cross-section of the Scorer parameter ( $\text{km}^{-1}$ ) at  $1^\circ$  of latitude, between the surface and approximately 55 km in the beginning of afternoon. Contours represent the potential temperature. Right : domain-averaged vertical profile of the Scorer parameter at the same local time. Direction of the zonal is indicated with the white arrow.

230 wavelength is approximately 150 km, one order of magnitude larger to typical Earth's trapped  
 231 lee waves (Ralph et al., 1997). Large horizontal wavelengths are known to increase the vertical  
 232 leakage of trapped lee waves (Durrán et al., 2015). The horizontal wavelength of the trapped  
 233 lee waves generated above Atla Regio is consistent with VEx/VIRTIS nightside measurements  
 234 in the low cloud region (Peralta et al., 2008), whereas the trapped lee waves generated above  
 235 Aphrodite Terra and Beta Regio are larger by at least a factor of 2.

236 One of the main questions about the vertical propagation of bow-shaped waves on Venus  
 237 (and the reason why they can be detected at high altitudes by Akatsuki) is how the waves can  
 238 propagate through two (near-neutral) mixed layers. A similar configuration can be found in  
 239 the Earth's oceans where the seasonal and main thermoclines can be separated by a relatively  
 240 weakly stratified region (Eckart, 1961). An analogous case in the Earth atmosphere is the  
 241 propagation of gravity waves from the troposphere to the upper mesosphere and lower ther-  
 242 mosphere, tunneling through an evanescent region (Walterscheid et al., 2001). Gravity waves  
 243 are evanescent in neutral-stability layers, in which the energy of the waves decrease exponen-  
 244 tially with altitude. However, if the vertical extension of the neutral region is not too large,  
 245 it is possible for a significant fraction of the incident wave energy to pass through the neutral  
 246 stability region. By analogy with quantum mechanics, some transfer of energy is possible via  
 247 wave tunneling (Sutherland and Yewchuk, 2004) through a neutral barrier. To quantify the

248 energy transmitted from a region with a Brunt-Väisälä frequency  $N$  through a barrier of zero  
 249 Brunt-Väisälä frequency, the transmission  $T$  is defined as

$$T = \left[ 1 + \left( \frac{\sinh(k_x H)}{\sin 2\theta} \right)^2 \right]^{-1} \quad (1)$$

250 according to Sutherland and Yewchuk (2004), with  $k_x$  the horizontal wavenumber,  $H$  the height  
 251 of the barrier and  $\theta$  equals to  $\cos(\omega/N)$  with  $\omega$  the frequency of the wave.

252 For the first barrier, between 18 and 35 km, the horizontal wavelength is of the order of  
 253 200 km,  $H$  is equal to 17 km,  $\omega$  to  $3.4 \cdot 10^{-4} \text{ s}^{-1}$  and  $N$  to  $2.3 \cdot 10^{-3} \text{ s}^{-1}$ . For the second barrier,  
 254 the convective layer between 48 and 52 km, the horizontal wavelength is of the order of 200 km,  
 255  $H$  is equal to 4 km,  $\omega$  to  $1.3 \cdot 10^{-3} \text{ s}^{-1}$  and  $N$  to  $9.0 \cdot 10^{-3} \text{ s}^{-1}$ . With these values, the transmission  
 256  $T$  for the first layer is equal to 21 % and to 84 % for the second barrier. The cloud convective  
 257 layer depth is closer to 10 km (Tellmann et al., 2009), with that value the transmission drops  
 258 to about 45 %.

259 We conclude that, despite the presence of the two neutral-stability layers in the atmosphere  
 260 of Venus, there is a significant energy transmission through these two barriers. In other words,  
 261 this tunneling effect enables the orographic gravity waves to propagate towards the top Venusian  
 262 clouds, where those waves are detected by Akatsuki (Fukuhara et al., 2017; Kouyama et al.,  
 263 2017).

264 The mixed layer in the deep atmosphere between altitudes 18 and 35 km (Schubert et al.,  
 265 1980) is the thickest of the two barriers, and is the one that affects the most the vertically-  
 266 propagating wave, since a fifth of the wave energy makes it through this mixed layer. Our  
 267 simulations thereby provide insights into the mechanisms responsible for the propagation of the  
 268 bow-shape perturbation from the surface to 35 km, left unexplained by the approach based on  
 269 gravity-wave parameterization used in Navarro et al. (2018). Conversely, the cloud convective  
 270 layer between altitudes 48 and 52 km, thinner than the deep atmosphere mixed layer but with  
 271 strong vertical plumes (Lefèvre et al., 2018), plays a minor role and does not affect substantially  
 272 the wave vertical propagation (more than 80% of the incoming wave energy is propagating  
 273 through this mixed layer). These two neutral barriers would also be an obstacle to the vertical  
 274 propagation of sound waves (Martire et al., 2018) induced by putative Venusian seismic activity,  
 275 possibly detectable via infrasound sensors on balloons (Krishnamoorthy et al., 2018) or through

276 airglow radiation perturbation (Didion et al., 2018).

277 Saturation of a wave occurs either through critical levels (when the speed of the background  
 278 horizontal flow is equal to the wave phase speed) or wave breaking through convective instability.  
 279 To quantify this probability, the saturation index  $S$  of Hauchecorne et al. (1987) is used

$$S = \sqrt{\frac{F_0 N}{\rho k_x |\bar{u} - c|^3}} \quad (2)$$

280 where  $F_0$  is the vertical momentum flux,  $N$  is the Brunt-Väisälä frequency,  $\rho$  the density of the  
 281 atmosphere,  $k_x$  the horizontal wave number  $c$  the phase speed of the wave, and  $\bar{\cdot}$  the average  
 282 over our chosen mesoscale domain (thought to represent the large-scale component). When  $S$   
 283 reaches values close to 1 the wave may be likely to break through critical level or saturation. For  
 284 the case of Venus, the main wave is orographic and stationary with respect to the surface, with  
 285 a phase speed  $c$  equals to zero. The zonal wind speed is constantly increasing up to roughly  
 286 80 km of altitude, thus  $|\bar{u} - c| \gg 0$  and superior to  $F_0 N / k_x$ , which causes  $S$  to be much smaller  
 287 than 1, even for the largest values of stability  $N$ . The saturation index only reaches values up  
 288 to  $1 \cdot 10^{-1}$ , and most often values of  $1 \cdot 10^{-2}$ , thus the probability of saturation of the mountain  
 289 waves is low.

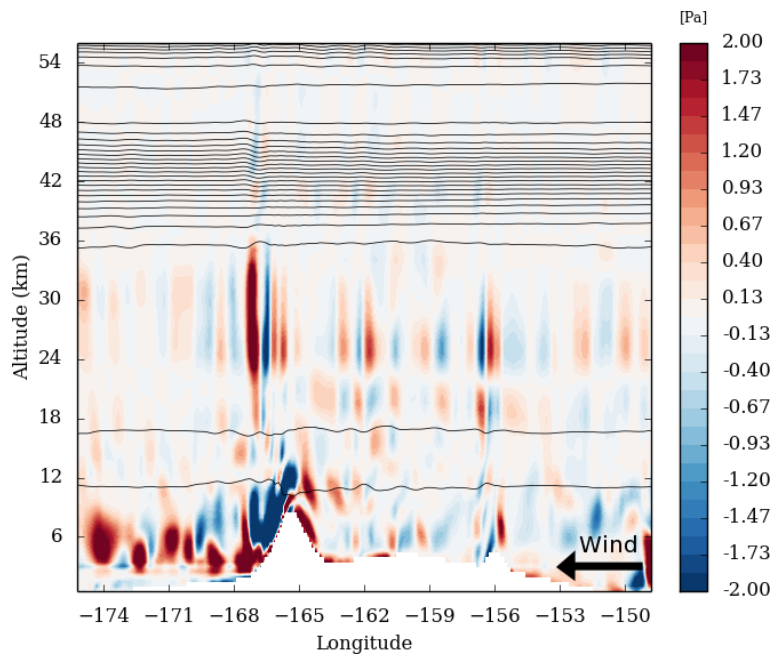


Figure 5: Instantaneous vertical cross-section of the quantity  $-\rho u'w'$  (Pa) at  $1^\circ$  of latitude, between the surface and approximately 55 km in the beginning of afternoon. Contours represent the potential temperature. Direction of the zonal is indicated with the black arrow.

290 The propagation of the waves into the two mixed layers engender a dissipation of the waves:  
 291 the part of the energy that is not transferred upward through tunneling effect is deposited in  
 292 those mixed layers. This deposition of momentum can be quantified by calculating the vertical  
 293 component of the Eliassen-Palm momentum flux (Andrews, 1987) that is equal to  $-\overline{\rho u' w'}$  where  
 294  $\rho$  is the density and  $u'$  and  $w'$  the perturbations of the zonal wind  $u$  and vertical wind  $w$  (with  
 295 respect to the mean defined as  $\bar{\cdot}$ ). A positive value of  $u'$  represents an eastward velocity pertur-  
 296 bation and a positive value of  $w'$  represents an upward velocity perturbation and therefore a  
 297 positive value of  $-\overline{\rho u' w'}$  represents an upward transport of westward momentum. The quantity  
 298 displayed at Figure 5 is an instantaneous snapshot of the quantity  $-\overline{\rho u' w'}$ . This flux is around  
 299 10 times larger than the strongest measured on Earth in Antarctica (Jewtoukoff et al., 2015).  
 300 This is partly due to the fact that the density of the Venusian atmosphere is 65 times larger than  
 301 the Earth atmosphere. Momentum is transported by the waves near the surface (where density  
 302 is the highest), but also in the mixed layer around 30 km of altitude (where wave perturbations  
 303 are strong). The trapped lee waves also transport momentum horizontally, while the vertical  
 304 transport of momentum due to leakage is negligible. The momentum flux  $-\overline{\rho u' w'}$  is the drag  
 305 coefficient (Smith, 1979) calculated in sub-grid-scale parameterizations, such as the one used in  
 306 the GCM runs by Navarro et al. (2018), who employed the orographic-wave parameterization  
 307 of Lott and Miller (1997). In this parameterization, Navarro et al. (2018) had to adopt values  
 308 of 2 Pa for the threshold of the gravity wave mountain stress, and 35 km for the initial altitude  
 309 of deposition of this mountain stress, to reproduce the bow-shape waves observed by Akatsuki.  
 310 Our mesoscale modeling of the Venusian mountain waves, in which the propagation of gravity  
 311 waves is resolved from the surface to 100 km, therefore validates the assumption of the Navarro  
 312 et al. (2018) study, both for the amplitude of the stress and the altitude of the stress deposition  
 313 (a maximum of momentum is deposited by the waves at this altitude), and shows that they are  
 314 physically based.

315 What we describe here for Alta Regio extends to the other regions considered in this study.  
 316 Figure 6 shows the vertical cross-section of the vertical wind, Scorer parameter and momentum  
 317 flux for Aphrodite Terra (left) and Beta Regio (right). As for Atla Regio, the major terrain  
 318 elevations generate gravity waves and the flow is in a non-linear regime. The two layers of  
 319 low static stability are also present in those regions, which entails the generation of trapped

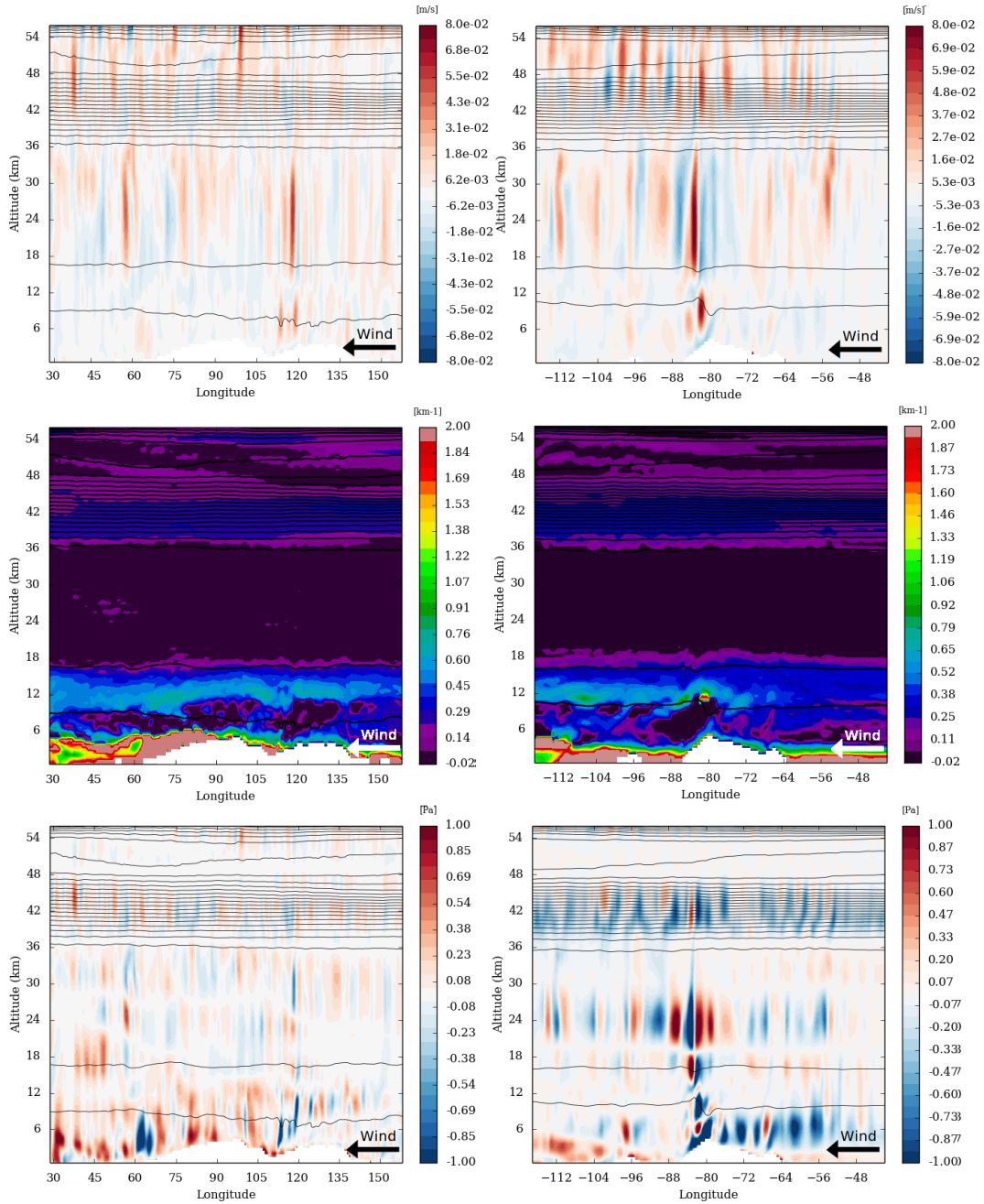


Figure 6: Vertical cross-section of the vertical wind ( $\text{m s}^{-1}$ ) on top panel and vertical cross-section of the Scorer parameter ( $\text{km}^{-1}$ ) on middle panel and vertical cross-section of vertical momentum flux (Pa) on bottom panel for Aphrodite Terra on left column and Beta Regio on right column. Contour is potential temperature. Direction of the zonal is indicated with arrows.

320 lee waves. The amplitude of the vertical wind is smaller than in the Atla case due to lower  
 321 slopes. Thus, the amplitude of the momentum flux is of the same order of magnitude, albeit  
 322 slightly smaller. The vertical extent of the two mixed layers is very similar to Atla Region, in  
 323 both the Aphrodite Terra and Beta Regio cases, but the horizontal wavelengths are slightly  
 324 larger and resulting in a small decrease of the tunneling transmission  $T$  but in the same order  
 325 of magnitude, about 20 % for the first barrier (18-35 km altitudes) and 80 % for the second

326 barrier (48-52 km altitudes).

## 327 4 Aphrodite Terra

328 Figure 7 shows the selected domain for Aphrodite Terra with a resolution of 40 km, Ovda Terra  
329 is visible between 60 and 100° of longitude where the main bow-shape gravity waves have been  
330 observed by *Akatsuki*.

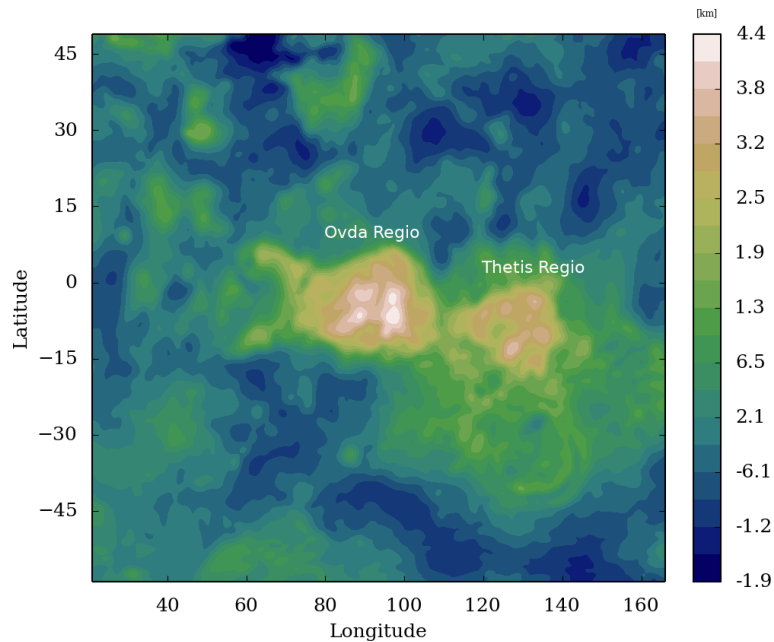


Figure 7: Elevation map (km) of the selected domain for Aphrodite Terra with a resolution of 40 km.

331 *Akatsuki* LIR measurements consist of a mean of the brightness temperature between 8  
332 and 12  $\mu\text{m}$  to which is applied a high-pass filter (Fukuhara et al., 2017; Kouyama et al., 2017;  
333 Taguchi et al., 2007). To be able to compare the outputs of the model with the observations,  
334 we have to analyze in the model the deformation of the cloud top by the gravity waves. We  
335 consider that the characteristic timescale of the wave propagation is smaller than the radiative  
336 timescale, so that the deformation is adiabatic; it follows that potential temperature can be  
337 used as a tracer for the deformation of the cloud top by waves. By choosing one value for  
338 potential temperature to set a relevant material surface, a corresponding temperature map  
339 can be reconstructed, as well as a map of the corresponding altitude. To compare the waves  
340 resolved by the model to the ones observed by the *Akatsuki* spacecraft, anomaly temperature  
341 is calculated for potential temperature surface between 55 and 80 km, corresponding to a range

342 of potential temperature from 800 to 1300 K, with a gaussian weighting function mimicking  
 343 the LIR’s weight function. Then the temperature perturbations are vertically averaged. A  
 344 high-pass gaussian filter is applied to filter out the global dynamics component (similarly to  
 345 what is done in the maps derived from Akatsuki observations Fukuhara et al., 2017; Kouyama  
 346 et al., 2017), and a low-pass gaussian filter is also applied to remove the small-scale features  
 347 smaller than a few tens of kilometer (i.e. approaching the effective horizontal resolution of our  
 348 mesoscale simulations for Venus). Figure 9 presents the associated residuals of the temperature  
 349 anomaly after filtering. In the end, temperature maps such as Figure 8 obtained from our  
 350 mesoscale modeling by this method can be directly compared to the *Akatsuki* observations in  
 351 Fukuhara et al. (2017) and Kouyama et al. (2017).

#### 352 4.1 Resulting bow-shape wave

353 Figure 8 shows the temperature anomaly modeled by our LMD Venus mesoscale model at  
 354 the top of the cloud above Aphrodite Terra, close to the evening terminator. Cyan contours  
 355 show the topography. The topography-induced perturbation at the top of the cloud expands  
 356 from -40 to 40° of latitude with a bow-shaped morphology, resembling the measured signal by  
 357 Akatsuki. Above Ovda Terra, the positive and negative anomaly measured by LIR (Fukuhara  
 358 et al., 2017) is reproduced with a similar amplitude,  $\pm 2$  K. Outside of the bow-shape wave  
 359 small-scale waves features are visible, similar structures are visible in the UV images (Kitahara  
 360 et al., 2019). This orographic gravity wave induces a deformation of about 600 m of the cloud  
 361 top altitude. Temperature anomalies induced by mountain waves are visible in the middle  
 362 and upper cloud layers whereas no temperature anomaly is discernible in the lower cloud layer  
 363 due to the presence of the convective layer. This could explain the fact that VEx/VIRTIS  
 364 measured on the nightside stationary waves in the upper cloud layer but not in the lower cloud  
 365 layer (Peralta et al., 2017).

366 The divergence of the momentum flux indicates the acceleration (or equivalently the force  
 367 per unit mass) caused by gravity waves on the mean flow when they break or encounter a  
 368 critical level (Frits and Alexander, 2003). This acceleration can be calculated by the equation

$$\frac{\partial \bar{u}}{\partial t} = \frac{1}{\rho} \frac{\partial}{\partial z} \overline{\rho u' w'} \quad (3)$$

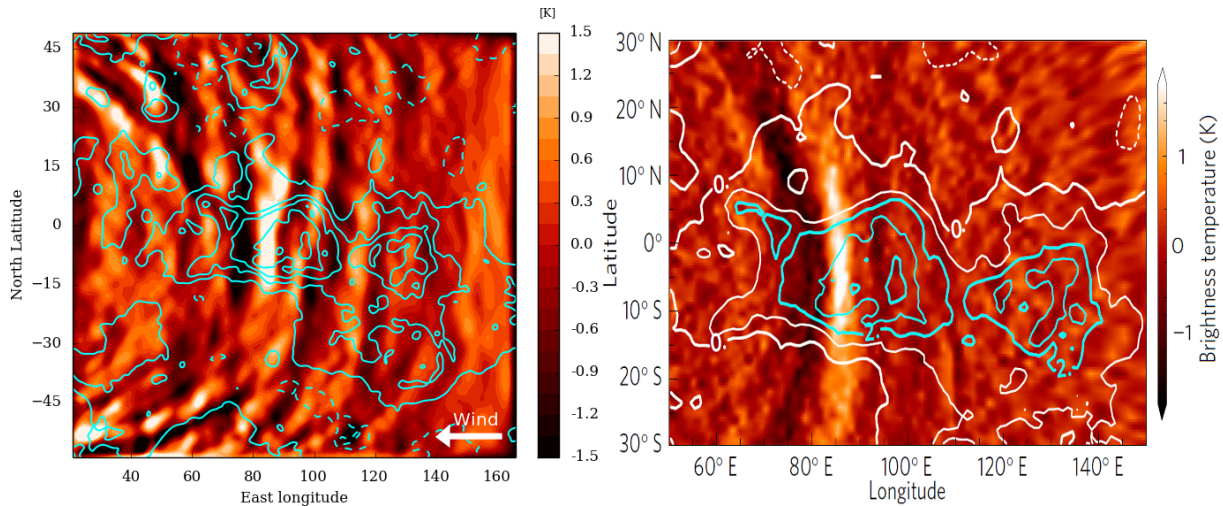


Figure 8: Left : Temperature anomaly (K) at the top of the cloud in late afternoon. Cyan contours show the topography every 1 km (Fig. 7). Direction of the zonal is indicated with the white arrow. Right : Temperature anomaly (K) at the top of the cloud observed by *Akatsuki*/LIR adapted from Fukuhara et al. (2017). White and cyan lines are topography.

369 Observations indicate that around cloud top, for altitudes above 67 km, there is a deceleration  
 370 of the zonal wind.

371 Bertaux et al. (2016) calculated that an deceleration of  $13 \text{ m s}^{-1}$  per Venus day was nec-  
 372 essary to explain longitudinal shift of zonal wind patterns. The *Akatsuki* spacecraft measured  
 373 a longitudinal variability of the zonal wind around cloud top between  $4$  and  $12 \text{ m s}^{-1}$  at the  
 374 Equator (Horinouchi et al., 2018) with no correlation with topography. The deceleration in-  
 375 duced by the bow-shape waves resolved by our LMD Venus mesoscale model, integrated over a  
 376 Venus day, reaches values around  $3 \text{ m s}^{-1}$  consistent with the lower range of estimates based on  
 377 *Akatsuki* measurements. This value is smaller than the computations in Bertaux et al. (2016)  
 378 (and the higher range of Akatsuki estimates), which tends to indicate that the deceleration of  
 379 zonal wind at cloud top cannot be explained solely by the impact of bow-shaped gravity waves.

## 380 4.2 Variability with local time

381 The temperature anomaly of 2 K at the top of the cloud shown in Figure8 is the maximum  
 382 signal obtained. This wave is visible in the model with an amplitude superior to 0.5 K for about  
 383 10 Earth days, against approximately 14 Earth days for the *Akatsuki* observation (between 14h  
 384 et 17h in local time). Figure 10 shows Temperature anomaly (K) at the top of the cloud  
 385 3.5 Earth day after Figure 8, the bow-shape wave above Ovda Regio is still visible but with  
 386 smaller amplitude meanwhile a bow-shape wave above Thetis Regio is discernible similarly to

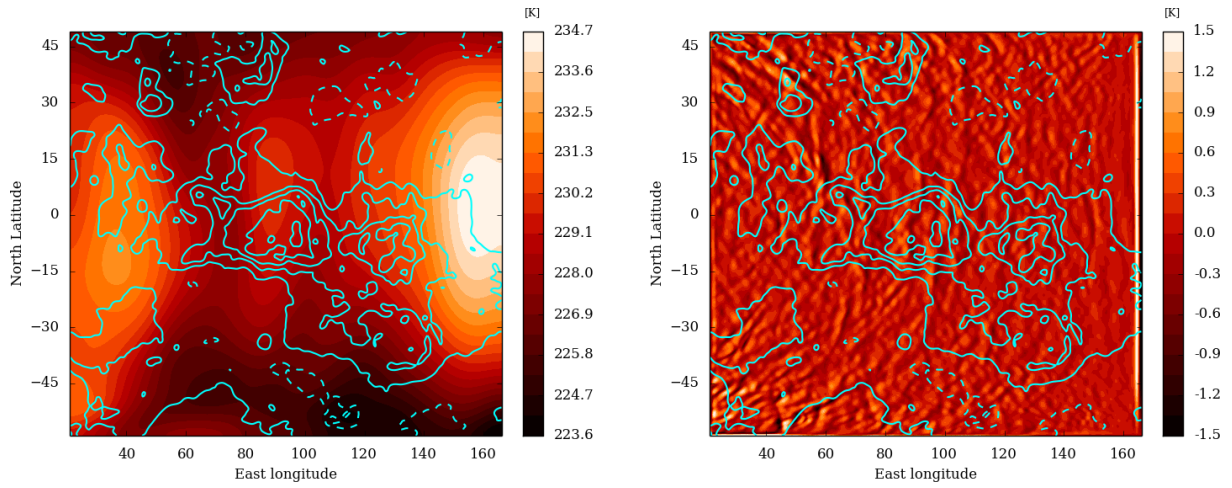


Figure 9: The scene displayed in this figure is similar to Figure 8. The low-pass filter signal (left panel) representing the global dynamics component has a maximum around  $160^\circ$  of longitude, where the local time is close to noon. The high-pass filter signal (right panel) shows small-scale wave features not visible on Akatsuki LIR images (Kouyama et al., 2017) but discernible on UV images (Kitahara et al., 2019). Cyan contours show the topography (Fig. 7).

387 LIR observations (Kouyama et al., 2017).

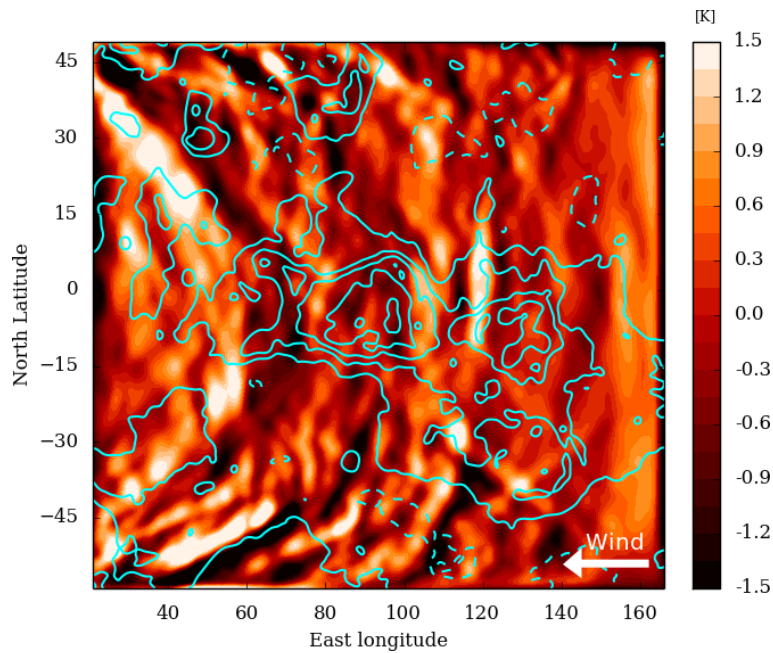


Figure 10: Temperature anomaly (K) at the top of the cloud 3.5 Earth day after Figure 8. Cyan contours show the topography every 1 km (Fig. 7). Direction of the zonal wind is indicated with the white arrow.

388 Additional simulations were performed at midnight and noon to study the variability of  
 389 the waves at cloud top. At midnight, no significant wave (amplitude superior to 0.5 K) is  
 390 observed. At noon, transient waves are visible with amplitude smaller than 1 K. The variation  
 391 of the surface wind along the day, about  $1 \text{ m s}^{-1}$  above Ovda Terra, is too small to explain  
 392 this variability. This indicates that the near-surface conditions responsible for the emission of

393 orographic gravity wave are not changing significantly along the day. The diurnal variability of  
 394 conditions responsible for the propagation of orographic gravity waves towards high altitude is  
 395 more likely to explain the observed variability with local time of the bow-shaped waves.

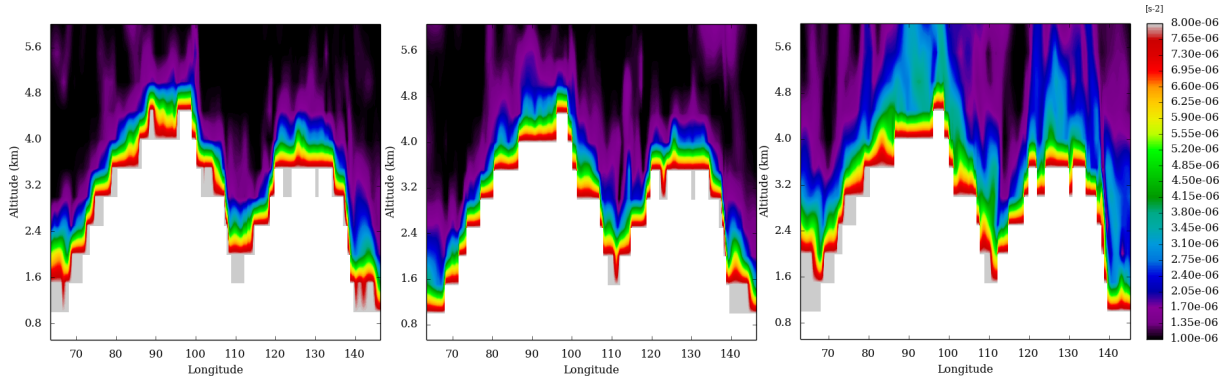


Figure 11: Square of the Brunt-Väisälä frequency ( $s^{-2}$ ) of the atmosphere from the surface to 6 km for midnight (left), noon (middle) and late afternoon (right) for Aphrodite Terra.

396 Figure 11 shows the square of the Brunt-Väisälä frequency ( $s^{-2}$ ) above Aphrodite Terra be-  
 397 tween the surface and 6 km of altitude for the three local times : midnight (left), noon (center),  
 398 late afternoon (right). Close to the surface the atmosphere is very stable, but this stability is  
 399 decreasing with altitude by an order of magnitude. This decrease is different depending on the  
 400 local time. At midnight, the static stability shows a strong gradient with altitude. At noon this  
 401 gradient is smaller, and even smaller in late afternoon with constant value of  $3 \times 10^{-6} s^{-2}$  over  
 402 more than one kilometer. This slow decrease has consequences over the Scorer parameter, it is  
 403 superior to 2 km in the first 8 km in late afternoon, against only in the first 3 km at other local  
 404 times. This strong value of the Scorer parameter over several kilometers favors the vertical  
 405 propagation of the gravity waves. This is correlated to the vertical momentum flux: at night  
 406 the amplitude is about  $10^{-2}$  Pa, while it is about  $10^{-1}$  Pa at noon. Meanwhile, the very-near  
 407 surface (first hundred meters) stability is consistent with Navarro et al. (2018) study : the at-  
 408 mosphere is more stable at night, but with a smaller local-time variability. This behavior plays  
 409 a role in the generation of the waves, while the first 4 to 5 km affect the vertical propagation.

## 410 5 Atla and Beta Regios

### 411 5.1 Atla Regio

412 The left panel of Figure 12 shows the resulting temperature anomaly at cloud top for the Atla  
413 Terra (Fig. 2) simulation, using the same methodology as for Aphrodite Terra. Two waves are  
414 visible, one above the main mountain Maat Mons at  $-168^\circ$  of longitude and an other over Ozza  
415 Mons at  $-163^\circ$  of longitude. The amplitude of these waves is of about  $\pm 1.5$  K for the first one  
416 and about  $\pm 1$  K for the last one. The latitudinal expansion is about  $20^\circ$ .

417 LIR observed one main wave above Ozza Mons with an amplitude of  $\pm 2$  K and an extension  
418 of  $30^\circ$  as well as a second wave of a few degrees of latitude associated with Maat Mons (Kouyama  
419 et al., 2017). The main resolved wave is similar to the observations, with however sharp  
420 morphology that may be due to the abrupt terrain elevation; the second wave is more extended  
421 in latitude than in the observations. These waves are obtained for the beginning of the afternoon  
422 when the amplitude is maximum, whereas LIR observed the maximum in the middle of the  
423 afternoon. The waves resolved by the model are symmetrical to the obstacle, meaning the  
424 latitudinal expansion towards the north and towards the south is the same, while LIR observed  
425 non-symmetrical waves attributed to the complex terrain.

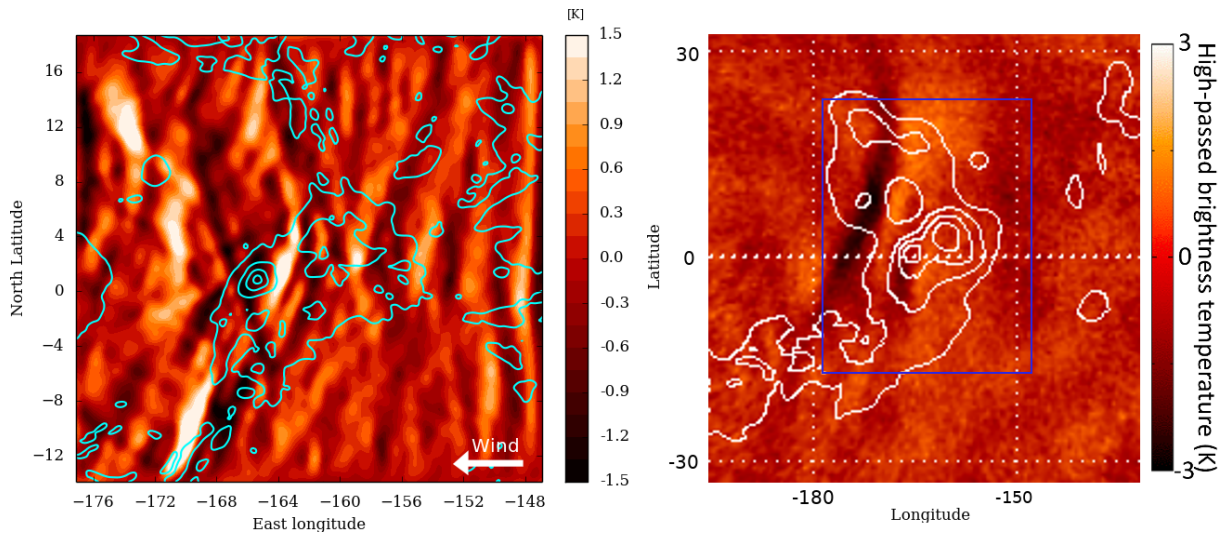


Figure 12: Left : Temperature anomaly (K) at the top of the cloud in the afternoon. Cyan contours show topography every 1.6 km (Fig. 2). Direction of the zonal is indicated with the white arrow. Right : Temperature anomaly (K) at the top of the cloud observed by *Akatsuki*/LIR adapted from Kouyama et al. (2017). White and cyan lines are the topography. Blue lines on the right panel show the mesoscale domain.

426 The generation and propagation of the waves are sensitive to the wind and stability condi-

427 tions close to the surface and therefore to physical representations within the model, for example  
 428 the radiative transfer. The Venus GCM may not represent fully accurate surface conditions,  
 429 especially for winds, as well as their evolution with time. These differences could explain that,  
 430 while the overall bow-shaped wave is reproduced in our Venus mesoscale simulation, some  
 431 discrepancies with the observations do exist.

## 432 5.2 Beta Regio

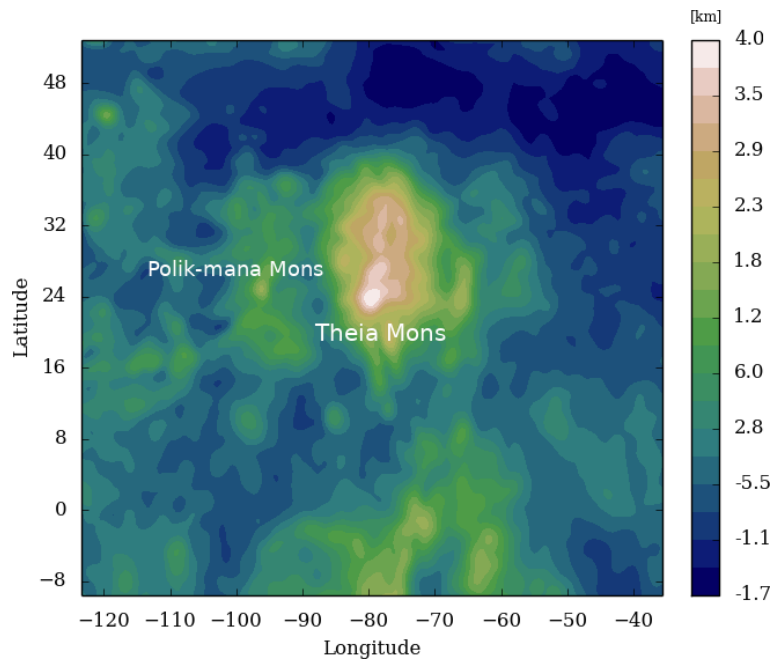


Figure 13: Elevation map (km) of the selected domain for Beta Regio with a resolution of 30 km.

433 Figure 13 shows the selected domain Beta Regio with a resolution of 30 km, where the  
 434 highest elevation point is Theia Mons at  $-80^\circ$  of longitude and  $30^\circ$  of latitude. The temperature  
 435 anomaly at the top of the cloud is shown in left panel of Figure 14. Two distinct stationary  
 436 waves are visible, a main one above Theia Mons and a smaller one above Polik-mana Mons  
 437 at  $30^\circ$  of latitude and  $-100^\circ$ . These two waves are observed by *Akatsuki*/LIR with the same  
 438 relative size visible in the right panel of Figure 14. The bow shape is visible with an amplitude  
 439 of  $\pm 1.5$  K. A difference with previous anomaly is the asymmetry of the waves, which is more  
 440 developed towards the North. This non-symmetrical shape is not clearly observed above Beta  
 441 Regio in LIR images (Kouyama et al., 2017) but visible in UV images (Kitahara et al., 2019).  
 442 The latitude of the region,  $30^\circ$ , may have an impact on the morphology of the wave.

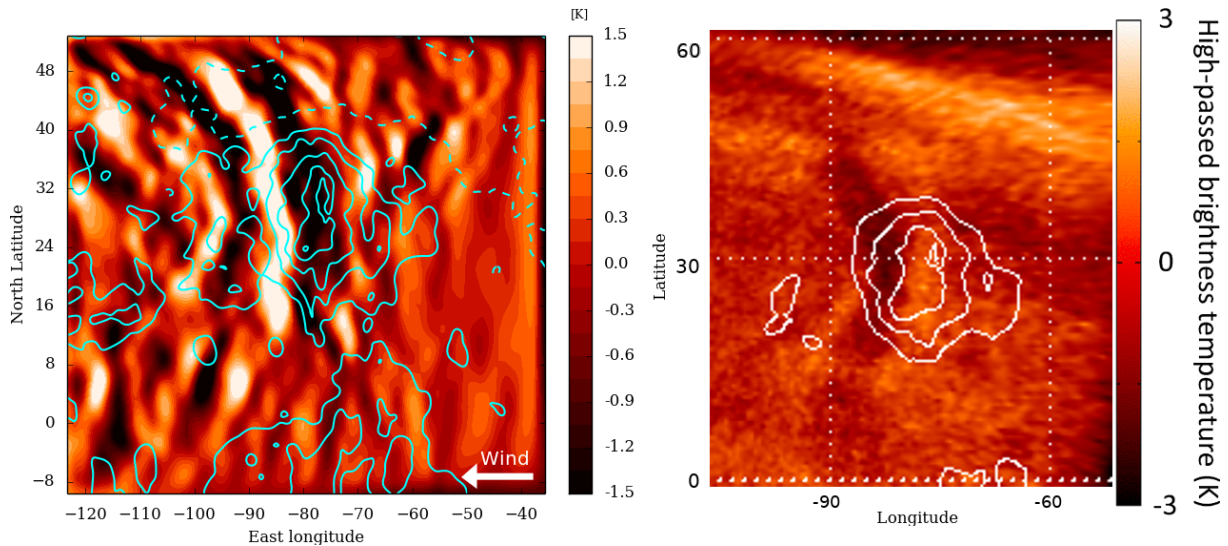


Figure 14: Left : Temperature anomaly (K) at the top of the cloud in late afternoon. Cyan contours show topography every 850 m (Fig. 13). Direction of the zonal is indicated with the white arrow. Right : Temperature anomaly (K) at the top of the cloud observed by *Akatsuki*/LIR adapted from Kouyama et al. (2017). White and cyan lines are the topography.

## 443 6 Polar regions

444 The highest topographical point is Maxwell Montes peaking at approximately 10 km at 65°N  
 445 latitude. This is a site of interest in orographic gravity waves on Venus. Unfortunately, the orbit  
 446 of *Akatsuki* is equatorial and therefore the spacecraft cannot observe the atmospheric activity  
 447 in polar regions; *Venus Express* had an elliptical polar orbit suited to study the south pole,  
 448 but not the north pole. To study the generation of orographic waves above Maxwell Montes,  
 449 and more generally in the polar regions, mesoscale simulations can be performed with polar  
 450 stereographic projection over both poles for comparison. The chosen domains for North (right)  
 451 and South pole (left) are shown in Figure 15, with a resolution of 40 km for both domains.  
 452 Recent improvements of the IPSL Venus GCM used to provide initial and boundary for our  
 453 Venus mesoscale model, especially in the polar regions, ensure a correct large-scale forcing with  
 454 the inclusion of the cold collar and mid-to-high-latitude jets (Garate-Lopez and Lebonnois,  
 455 2018).

456 Figure 16 shows the temperature anomaly at the top of the cloud. Transient waves rotating  
 457 around the planet are visible in both polar regions down to  $\pm 60^\circ$  with an amplitude larger  
 458 than 5 K. The wave structure and amplitude are similar for both regions meaning that the  
 459 underlying topography has hardly any impact on the temperature anomaly at cloud top. These  
 460 waves resemble the planetary streak structure visible in the *Akatsuki*/IR2 images (Limaye et al.,

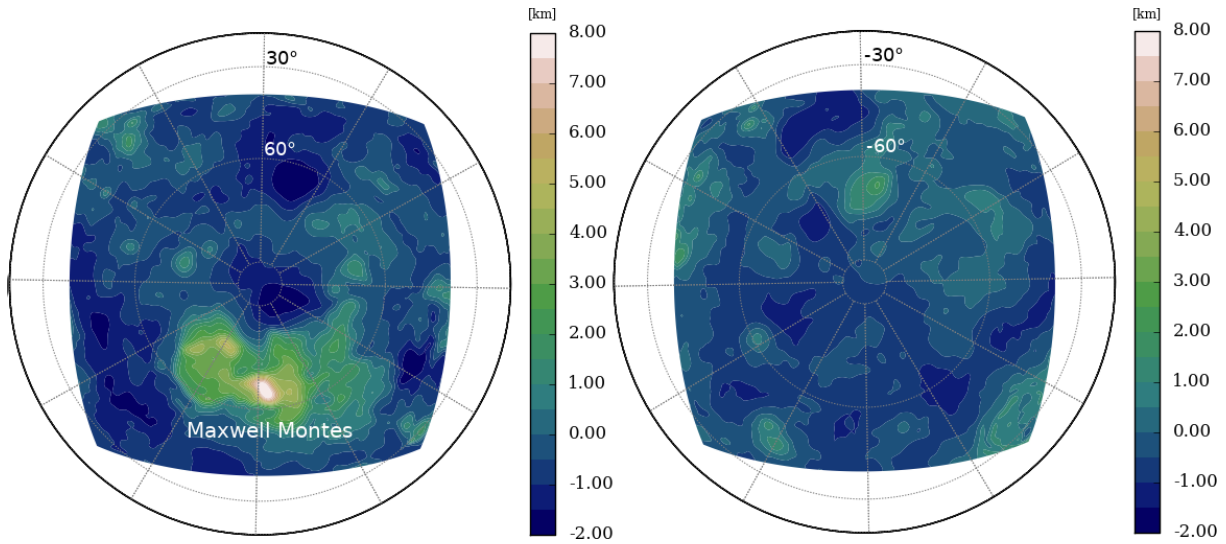


Figure 15: Elevation map (km) of the selected domains for the North pole (left) and the South pole (right) regions with a resolution of 40 km.

461 2018) and reproduced with GCM modelling (Kashimura et al., 2019).

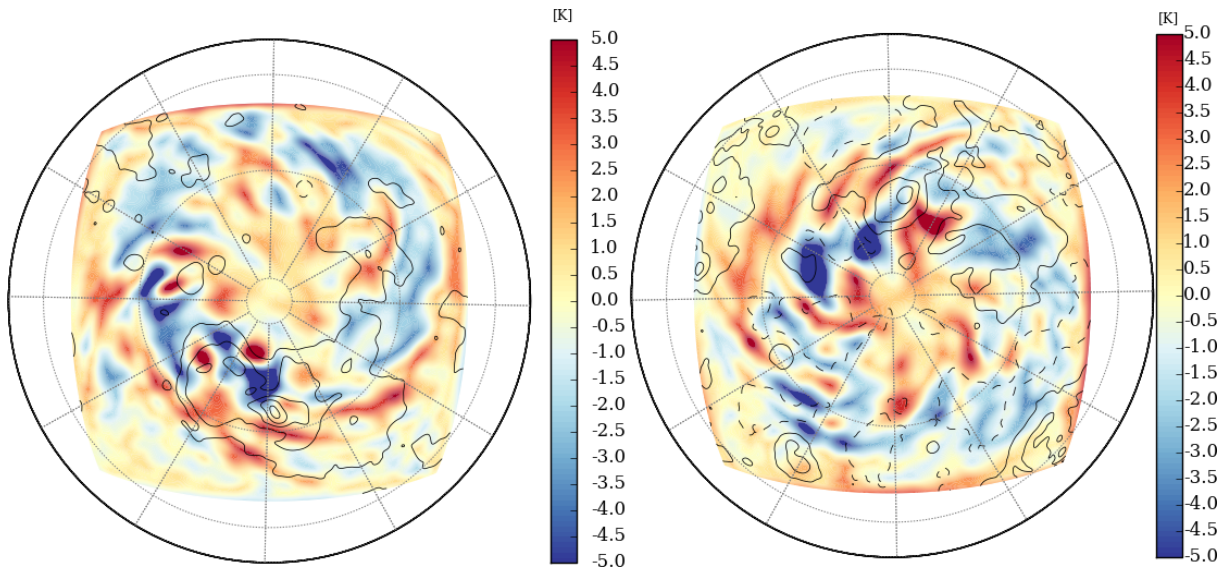


Figure 16: Temperature anomaly (K) at the top of the cloud for the North pole (left) and the South pole (right) regions. Black contours show topography every 1.6 km for the North pole and every 800 m for the South pole (Fig. 15).

462 However, mountain waves are generated by Ishtar Terra. The vertical velocity above  
 463 Maxwell Montes is shown in Figure 17. These waves propagate vertically and encounter the two  
 464 low static stability regions but do not yield any stationary temperature anomaly at cloud-top.  
 465 The two barriers are thicker than that for the other cases, 22 against 17 km for the first one and  
 466 7 against 4 km for the convective layer. The increase of the cloud convective layer thickness is  
 467 expected from observations (Tellmann et al., 2009; Haus et al., 2014) and modelling (Imamura  
 468 et al., 2014; Lefèvre et al., 2018), yet few observations have been performed regarding the mixed

469 layer in the deep atmosphere, especially regarding the variation of depth with latitude and local  
 470 time. Only the temperature profiles of the Pioneer Venus probes (Seiff et al., 1980) would be  
 471 available for such an analysis.

472 As is discussed in section 3, the two mixed layers in the lower atmosphere of Venus act  
 473 as barriers for gravity waves. For the first barrier (altitudes 15-35 km) with  $\omega$  equals to  
 474  $2.3 \cdot 10^{-4} \text{ s}^{-1}$ ,  $N$  to  $1.6 \cdot 10^{-3} \text{ s}^{-1}$  and a horizontal wavelength of 200 km, the transmission is  
 475 only of 13 %. For the second barrier (altitudes 45-55 km) with  $\omega$  equals to  $1.1 \cdot 10^{-3} \text{ s}^{-1}$ ,  $N$   
 476 to  $7.5 \cdot 10^{-3} \text{ s}^{-1}$  and the same horizontal wavelength, the transmission drops to 63 %. The  
 477 orographic wave is therefore strongly affected by propagating through the two mixed layers  
 478 (especially the lowermost one). As a result, the cloud top temperature anomaly induced by  
 479 gravity waves in polar regions is smaller than the ones in the equatorial regions, and negligible  
 480 against the cloud top transient waves. The large amplitude of the transient polar waves could  
 481 prevent the visibility of the comparatively-low stationary gravity-wave perturbation and explain  
 482 the non-observation of stationary waves in *Venus Express*/VIRTIS data at latitudes above  $65^\circ\text{S}$   
 483 (Peralta et al., 2017). The saturation of the waves is here again unlikely, so contrary to mixed  
 484 layers acting as efficient barriers, it does not seem a plausible mechanism to explain the lack of  
 485 observed gravity-wave stationary wave in polar regions.

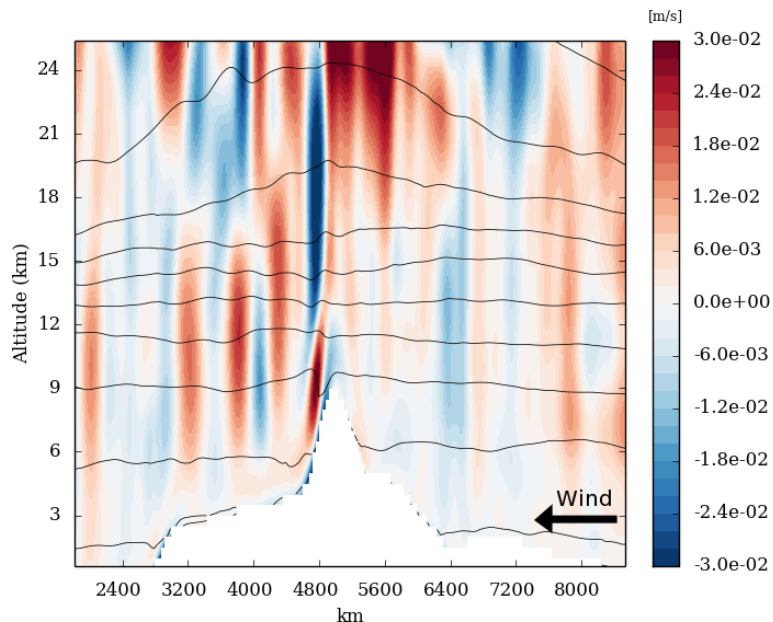


Figure 17: Vertical cross-section of the vertical wind ( $\text{m s}^{-1}$ ) between the surface and approximately 55 km. Please note that the cross-section is not at a constant latitude. Contours represent the potential temperature.

## 486 7 Conclusion

487 We present here the first mesoscale model applied to Venus, in which the WRF dynamical core  
488 is coupled the set of physical parameterizations developed for the IPSL Venus GCM, notably  
489 the radiative transfer to compute the solar and IR rates during the simulations. Focusing  
490 on three areas of interest, the model resolves bow shape stationary waves with an amplitude  
491 around 1.5 K. The lifetime of the waves is about ten Earth days. The maximum amplitude  
492 of the waves are in the afternoon, close to the terminator for Aphrodite Terra, earlier for the  
493 two another cases. The characteristics and morphology of the waves observed by Akatsuki are  
494 overall well reproduced by the model.

495 These gravity waves are generated by the large-scale flow forced to go over Venus' moun-  
496 tains. Propagating vertically, those orographic gravity waves encounter two layers of low static  
497 stability, the mixed layer (20-32 km) and the convective layer (47-55 km) where some energy is  
498 transmitted through the layers via tunneling phenomenon. The deep mixed layer is the most  
499 critical barrier for the wave, and comprehensive studies about the variability of this layer with  
500 latitude and local time would improve the understanding of the propagation of the mountain  
501 waves. The presence of these two layers generate trapped lee waves that propagate horizontally  
502 with some vertical leakage due to their large horizontal wavelength.

503 The variations of the wave characteristics with local time is imputed to the stability of the  
504 atmosphere close to the surface. In the afternoon, the first 4-5 km of the atmosphere are more  
505 stable than at night or at noon, which favors the vertical propagation of the waves.

506 The waves extract momentum from the surface with a flux around 2 Pa and deposit this  
507 momentum at around 30 km. Values of parameters used in the orographic parameterization  
508 of Navarro et al. (2018) are therefore confirmed by our mesoscale model which resolves the  
509 emission and propagation of gravity waves from the surface to 100-km altitude. At the cloud  
510 top, the waves induce a deceleration of several meters per second over a Venus day, consistent  
511 with *Akatsuki* measurements.

512 Temperature anomalies above Venus' polar regions have also been explored with our mesoscale  
513 model. Transient waves rotating around the poles dominate the signal and the influence of the  
514 underlying topography is not noticeable. However, mountain waves are generated above Ishtar  
515 Terra – similarly to the waves generated by the other main mountains at lower latitudes. The

516 increase of the deep atmosphere neutral barrier thickness strongly affects the amplitude of the  
517 orographic waves, which become negligible against the circumpolar transient waves.

518 Surface wind and the two mixed layers are key factors for the generation and propagation of  
519 mountain waves, a sensitivity study should be therefore considered in future work. Such study  
520 will imply complex changes in the IPSL Venus GCM dynamics and cloud model.

521 Additional simulations may be performed in the future for several other regions of interest,  
522 like Pheobe Regio at the Equator with a complex bow-shape morphology witnessed by Akatsuki,  
523 Gula Mons and Bell Regio, two sharp mountains at respectively 20°N and 30°N and Imdr  
524 Regio composed of two sharp mountains at 45°S to investigate the influence of the impact of  
525 the morphology of the mountain, of the conditions near the surface and of the latitude on the  
526 temperature anomaly. Moreover the observations of stationary waves by VEx/VIRTIS in the  
527 southern hemisphere in the nightside with no apparent dependence with the local time (Peralta  
528 et al., 2017) are features to be investigated.

529 The LMD Mesoscale model for Venus based on WRF (Skamarock and Klemp, 2008) can  
530 be applied to the study of several other regional-scale phenomena, in particular slope winds,  
531 which were studied on Earth (Bromwich et al., 2001) and on Mars (Spiga et al., 2011), and  
532 play a key role in the vertical extension of the planetary boundary layer convection (Lebonnois  
533 et al., 2018) and local meteorology in general.

534 Implementation of the photochemistry and microphysics schemes developed at IPSL in the  
535 Venus mesoscale model is planned to investigate the influence of mountains on the chemistry  
536 and the cloud formation. Explosive volcanism could explain SO<sub>2</sub> anomaly at the top of the cloud  
537 (Esposito et al., 1988): the study of the vertical transport and the impact of the waves on SO<sub>2</sub>  
538 (Glaze, 1999) and other volatile like water (Airey et al., 2015) could constrain this phenomenon,  
539 especially above Atla Regio where hot spots have been observed with *Venus Express* suggesting  
540 active volcanism (Shalygin et al., 2015). Atmospheric variability on its own, without the need  
541 to invoke active volcanism, could also explain the observed SO<sub>2</sub> variability (Marcq et al., 2013).

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