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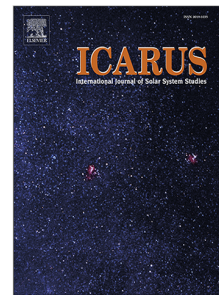
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main latex file

1 Radiative-equilibrium model of Jupiter's  
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## Abstract

Jupiter's upper troposphere and stratosphere are host to a rich dynamical and chemical activity. This modulates the thermal structure and distribution of trace species and aerosols, which, in turn, impact the atmospheric radiative budget and dynamics. In this paper, we present a computationally efficient 1-D seasonal radiative model, with convective adjustment, of Jupiter's atmosphere. Our model takes into account radiative forcings from the main hydrocarbons (methane, ethane, acetylene), ammonia, collision-induced absorption, several cloud and haze layers and an internal heat flux. We parametrize four tropospheric cloud and haze layers. Two of them (one tropospheric cloud near 800 mbar, one upper tropospheric haze, one stratospheric haze) are set to be uniform with latitude. On the contrary, we prescribe the spatial distribution of another UV-absorbing "polar" stratospheric haze comprising fractale aggregates based on published observational constraints, as their concentration varies significantly with latitude. We detail sensitivity studies of the equilibrium temperature profile to several parameters (hydrocarbon abundances, cloud particle sizes and optical depths, optical properties of the stratospheric polar haze, etc.). We then discuss the expected seasonal, vertical and meridional temperature variations in Jupiter's atmosphere and compare the modeled thermal structure to that derived from Cassini and ground-based thermal infrared observations.

We find that the equilibrium temperature in the 5–30 mbar pressure range

34 is very sensitive to the chosen stratospheric haze optical properties, sizes and  
 35 number of monomers. One of the three sets of optical properties tested yields  
 36 equilibrium temperatures that match well, to first order, the observed ones.  
 37 In this scenario, the polar haze significantly warms the lower stratosphere  
 38 (10–30 mbar) by up to 20K at latitudes 45–60°, and reproduces an observed  
 39 north-south asymmetry in stratospheric temperature. The polar haze also  
 40 acts to shorten significantly the radiative timescales, estimated by our model  
 41 to 100 (Earth) days at the 10-mbar level. At pressures lower than 3 mbar,  
 42 our modeled temperatures systematically underestimate the observed ones  
 43 by  $\sim 5$ K. This might suggest that other processes, such as dynamical heating  
 44 by wave breaking or by eddies, or a coupling with thermospheric circulation,  
 45 play an important role. An alternate possibility is that the uncertainty on  
 46 the abundance of hydrocarbons is responsible for this mismatch. In the  
 47 troposphere, we can only match the observed lack of meridional gradient of  
 48 temperature by varying the internal heat flux with latitude.

49 We then exploit knowledge of heating and cooling rates (using our ra-  
 50 diative seasonal model combined to observational constraints on the temper-  
 51 ature) to diagnose the residual-mean circulation in Jupiter’s stratosphere.  
 52 This is done under the assumption that the eddy heat flux convergence term  
 53 is negligible. In the Earth’s stratosphere, the residual-mean circulation ob-  
 54 tained with this method represents well, on a seasonal scale, the transport  
 55 of tracers in regions where wave breaking and dissipation are weak. How-  
 56 ever, on Jupiter, in the lower stratosphere (5–30 mbar), the residual-mean

57 circulation strongly depends on the assumed properties of the stratospheric  
58 haze. Our main conclusion is that it is crucial to improve our knowledge on  
59 the different radiative forcing terms (in particular regarding the stratospheric  
60 haze properties) to increase our confidence in the estimated circulation. By  
61 extension, this will also be crucial for future 3D GCM studies.

## 62 Highlights

- 63 • A seasonal radiative-convective model of Jupiter's atmosphere is pre-  
64 sented.
- 65 • Stratospheric polar haze greatly impact the equilibrium temperatures.
- 66 • We evaluate the residual-mean stratospheric circulations and discuss  
67 caveats.

## 1 Introduction

Jupiter's troposphere hosts a rich dynamical activity with strong, alternately eastward and westward zonal jets at low and mid-latitudes, many vortices in the polar regions (unveiled by Juno, Adriani et al., 2018), numerous planetary-scale and mesoscale waves, hotspots and disturbances (e.g., Choi et al., 2013). Jupiter's stratosphere is as dynamically active – if not more – than the troposphere, yet has received less attention comparatively to the large body of modeling work on the jovian tropospheric dynamics (Williams, 2003; Heimpel et al., 2005; Showman, 2007; Schneider and Liu, 2009; Young et al., 2019). The observed temperature field features numerous wave signatures (Li et al., 2006; Fletcher et al., 2017) and isolated disturbances, both in the tropics and in the auroral regions (Flasar et al., 2004; Sinclair et al., 2017). Furthermore, a large variability of stratospheric temperature from one Earth year to another is observed at Jupiter's equator (Fletcher et al., 2016). This is associated with the quasi-quadriennial oscillation (QJO), a periodic oscillation in zonal wind and temperature thought to result from wave-mean zonal flow interactions (Leovy et al., 1991; Orton et al., 1991; Flasar et al., 2004; Simon-Miller et al., 2007; Cosentino et al., 2017).

Stratospheric circulations are still poorly known, and are currently mostly deduced from the observation of anomalies in the distribution of trace species. For instance, in the middle stratosphere (1-10 hPa), the observed meridional distributions of ethane ( $\text{C}_2\text{H}_6$ ) and acetylene ( $\text{C}_2\text{H}_2$ ) – main by-products of

90 the methane photochemistry – are found to be at odds with the predictions  
 91 of one-dimensional (1-D) neutral photochemical models. While acetylene  
 92 is maximum at low latitudes, following the yearly-averaged insolation as is  
 93 expected from photochemistry, long-lived ethane increases towards the poles  
 94 (e.g., Nixon et al., 2007). Other puzzling observations are related to molecules  
 95 produced following comet Shoemaker-Levy 9 (SL 9) impact in Jupiter’s atmo-  
 96 sphere in 1994 and their subsequent migration (Moreno et al., 2003; Griffith  
 97 et al., 2004; Lellouch et al., 2006; Cavalié et al., 2017). One of the most  
 98 striking results is that, 6.5 years following the impact, HCN was found to be  
 99 efficiently mixed from the impact site (44°S) to northern mid-latitudes while  
 100 CO<sub>2</sub> was found to be greatly enhanced near the south pole (Lellouch et al.,  
 101 2006). In an attempt to explain the observed opposite distributions of C<sub>2</sub>H<sub>6</sub>  
 102 and C<sub>2</sub>H<sub>2</sub>, or HCN and CO<sub>2</sub>, several models including parameterizations of  
 103 meridional and vertical diffusion and advection have been proposed (e.g.,  
 104 Hue et al., 2018; Lellouch et al., 2006). Unfortunately, none could satisfac-  
 105 torily reproduce the observations. In short, there is currently no consistent  
 106 picture of Jupiter’s stratospheric circulations and how the distributions of  
 107 trace species are impacted by those circulations. The mechanism(s) forcing  
 108 the aforementioned stratospheric circulations are also unknown, in particu-  
 109 lar regarding the role of wave activity in the troposphere and stratosphere –  
 110 by analogy with the Brewer-Dobson circulation in the Earth’s stratosphere  
 111 (Butchart, 2014) – and that of radiative processes. In this paper, we focus on  
 112 the precise evaluation of radiative forcings with a 1-D radiative equilibrium



113 model while the study of wave forcing is devoted to future work.

114 Understanding in detail the radiative forcings in Jupiter's atmosphere is  
 115 also key to interpreting the observed thermal structure. At mid-latitudes  
 116 and near the 10-mbar level, a 5 to 10 K temperature contrast is reported  
 117 between the summer and winter hemispheres despite Jupiter's low obliquity  
 118 ( $3^\circ$ ) (Fletcher et al., 2016). An explanation was proposed by Zhang et al.  
 119 (2013a) who reported large radiative relaxation timescales near 10 mbar,  
 120 which could lead to a seasonal lag in the atmosphere's response to seasonal  
 121 forcing. However, their study only included forcing from gaseous compounds.  
 122 In a follow-up study, Zhang et al. (2015) highlighted the importance of ra-  
 123 diative forcing by stratospheric aerosols of auroral origin, which were found  
 124 to dominate the radiative heating at mid- and high- latitudes (instead of  
 125 methane). However, the impact of including these terms on the temperature  
 126 field, and its seasonal variations, was not studied.

127 Regarding the upper troposphere (100–500 hPa) and focusing on the  
 128 zonal-mean temperature, the cloudy equatorial zone is found to be 2–4K  
 129 colder than the warmer and clearer north and south equatorial belts, while  
 130 there is little meridional temperature contrast at mid-latitudes ( $30^\circ\text{N}$ – $70^\circ\text{N}$   
 131 and  $30^\circ\text{S}$ – $70^\circ\text{S}$ ) (Fletcher et al., 2016). These temperature variations are  
 132 supposedly linked to tropospheric circulations, however, the radiative contri-  
 133 bution from clouds and aerosols have not been studied quantitatively.

134 The aforementioned findings and open questions suggest that a complex  
 135 interplay of dynamical and chemical activity takes place in Jupiter's middle

136 atmosphere, modulating the thermal structure and the distribution of trace  
 137 species and aerosols, which in turn impact the radiative budget and dynam-  
 138 ics. All these observations and open questions motivate the development of a  
 139 Global Circulation Model (GCM) for Jupiter extending to the upper strato-  
 140 sphere. Such a model would eventually take into account three-dimensional  
 141 (3-D) dynamics, radiative forcings, photochemistry, cloud/aerosol micro-  
 142 physics and the couplings between them, including troposphere-stratosphere  
 143 interactions. Several attempts have been made in this direction (Yamazaki  
 144 et al., 2004; Schneider and Liu, 2009; Showman et al., 2019; Young et al.,  
 145 2019) illustrating the modeling complexity and high computational cost nec-  
 146 essary to address the questions opened by observations.

147 Our goal is to obtain a Jupiter GCM capable of combining radiative  
 148 transfer with high-resolution dynamics, akin to the approach we followed for  
 149 Saturn’s atmosphere (Guerlet et al., 2014; Spiga et al., 2020). In this pa-  
 150 per, we focus on the efficient parametrization of a radiative-convective model  
 151 in Jupiter’s upper troposphere and stratosphere. This model is to be later  
 152 coupled with a hydrodynamical solver to build a Jupiter GCM capable of  
 153 studying both tropospheric and stratospheric circulations. Apart from being  
 154 part of a GCM, such a 1-D radiative-convective model can be used to com-  
 155 pute radiative timescales (Kuroda et al., 2014; Zhang et al., 2013a) and, when  
 156 confronted with observations, can be a useful tool to diagnose whether to first  
 157 order radiative processes govern or not the thermal structure of the atmo-  
 158 sphere (e.g. Guerlet et al., 2014). Kuroda et al. (2014) have developed such

159 a radiative equilibrium model of Jupiter's stratosphere. They investigated  
 160 the sensitivity of the equilibrium temperature profiles to the assumed hydro-  
 161 carbon abundances and compared their results to a reference temperature  
 162 profile obtained near the equator by Galileo. However, Kuroda et al. (2014)  
 163 's model neglected the radiative impact of tropospheric and stratospheric  
 164 aerosols that are expected to play an important role in heat absorption and  
 165 redistribution. In this paper, we propose to refine the approach proposed by  
 166 Kuroda et al. (2014) by including the missing radiative contributions and to  
 167 extend the comparisons of our seasonal model to more recent observations.

168 Finally, knowledge of heating and cooling rates (diabatic forcings) can  
 169 also be exploited to diagnose the residual-mean circulation in the strato-  
 170 sphere, as was done by West et al. (1992) and Moreno and Sedano (1997).  
 171 This permits an estimate of the stratospheric circulation and transport of  
 172 tracers without building a GCM to resolve the dynamics, under the limiting  
 173 assumption that eddy heat flux is negligible compared to diabatic forcings.  
 174 In theory, this circulation can then be exploited to re-visit the interpreta-  
 175 tion of observed distribution of trace species, as was done by Friedson et al.  
 176 (1999) to address the spreading of dust following the impact of comet SL-  
 177 9. The studies by West et al. (1992) and Moreno and Sedano (1997) were  
 178 based on observations from the Voyager epoch, and an update of this type of  
 179 work based on Cassini-era observations and an up-to-date radiative transfer  
 180 model is needed. This is especially critical because, while West et al. (1992)  
 181 and Moreno and Sedano (1997) both agreed on the importance of including

182 heating by stratospheric aerosols, the circulations they obtained differ both  
 183 quantitatively and qualitatively.

184 In what follows, we describe a state-of-the art radiative-convective model  
 185 for Jupiter’s atmosphere (as part of a GCM to be presented in another paper),  
 186 present comprehensive comparisons to recent temperature observations and  
 187 exploit knowledge of the net radiative heating field to compute the residual-  
 188 mean circulation in Jupiter’s stratosphere. Section 2 describes our radiative  
 189 transfer model for Jupiter’s upper troposphere and stratosphere that includes  
 190 up-to-date spectroscopic parameters, an internal heat flux, radiative effects of  
 191 tropospheric clouds and aerosols as well as stratospheric aerosols comprising  
 192 fractal aggregates. We present the resulting thermal structure and compare it  
 193 with recently published ground-based and Cassini observations in Section 3.  
 194 We then detail our methodology to compute the residual-mean circulation in  
 195 Section 4 and discuss these results in Section 5.

## 196 **2 Jupiter radiative-convective model**

### 197 **2.1 Overall description**

198 Our Jupiter radiative-convective model is adapted from its Saturn counter-  
 199 part, described in detail in Guerlet et al. (2014). The two giant planets  
 200 Jupiter and Saturn share many characteristics and, as a result, the main  
 201 physical parametrizations are the same: a  $k$ -distribution model is used to  
 202 compute gaseous opacities (Goody and Yung, 1989), the radiative transfer

equations (including multiple scattering and Rayleigh scattering) are solved under a two-stream approximation, and a convective adjustment scheme relaxes – instantaneously – the temperature profile towards the adiabatic lapse rate when unstable lapse rates are encountered. An internal heat flux, set to  $7.48 \text{ W.m}^{-2}$  as determined by Li et al. (2018b), is also added as a radiative flux at the bottom of our model.

Jupiter’s diurnal cycle is neglected: a sensitivity test shows that the maximum amplitude of diurnal temperature variations is less than 0.1 K. Similarly, given the long radiative timescales in Jupiter’s atmosphere, heating and cooling rates are computed – and the temperature updated accordingly – every 10 jovian days. We take into account Jupiter’s small obliquity ( $3.13^\circ$ ) and the moderate eccentricity of its orbit (0.048) that is expected to play a role in the seasonal cycle. Jupiter’s perihelion occurs at a solar longitude of  $L_s=57^\circ$ , which is close to the summer solstice in the northern hemisphere (defined as  $L_s=90^\circ$ ,  $L_s$  being the heliocentric longitude of Jupiter counted from the northern spring equinox). Hence, northern summer is expected to be warmer than southern summer – at least in the stratosphere where radiative timescales are shorter than a season (Kuroda et al., 2014). If Jupiter’s seasonal forcing was dominated by eccentricity rather than obliquity, one could even expect to get warmer temperatures in southern “winter” ( $L_s=90^\circ$ ) than during southern “summer” ( $L_s=270^\circ$ ).

Apart from the orbital and planetary parameters, the magnitude of the internal heat flux and the absence of opaque rings, the main differences be-

226 tween Saturn and Jupiter radiative models relate to the gaseous composition  
 227 as well as cloud and haze properties, detailed below.

## 228 **2.2 Gaseous opacities and $k$ -distribution model**

229 Our Jupiter radiative model takes into account gaseous opacity from the  
 230 three main hydrocarbons: methane ( $\text{CH}_4$ ), ethane ( $\text{C}_2\text{H}_6$ ), acetylene ( $\text{C}_2\text{H}_2$ ),  
 231 along with collision-induced transitions by  $\text{H}_2$ - $\text{H}_2$  and  $\text{H}_2$ -He. Through their  
 232 infrared emissions, these molecules are the major stratospheric coolants,  
 233 while atmospheric heating is primarily due to absorption of visible and near-  
 234 infrared solar radiation by methane and aerosols. Furthermore, we also take  
 235 into account opacity from ammonia ( $\text{NH}_3$ ) that was previously neglected in  
 236 the Saturn model, as is justified in section 2.4.

237 As line-by-line calculations of absorption coefficients are too time-consuming  
 238 for the GCM runs we are aiming at, we use the correlated- $k$  method for the  
 239 computation of the atmospheric transmission at each time step. Correlated- $k$   
 240 coefficients are pre-tabulated offline on a 2D temperature-pressure grid com-  
 241 prising twelve temperatures points from 70 to 400K and nine pressure levels  
 242 from 10 to  $10^{-6}$  bar (one level every pressure decade, plus one level at 0.5  
 243 bar as ammonia varies rapidly with altitude in this region). To obtain these  
 244 tables, high-resolution absorption coefficient spectra  $k(\nu)$  are first computed  
 245 using the KSPECTRUM line-by-line model (Eymet et al., 2016) for a mix-  
 246 ture of gases ( $\text{CH}_4$ ,  $\text{C}_2\text{H}_6$ ,  $\text{C}_2\text{H}_2$ ,  $\text{NH}_3$ ) at each point of the  $(T, p)$  grid. The  
 247 gaseous abundance profiles used are detailed in section 2.4. A Voigt line

248 shape is assumed except for  $\text{CH}_4$ , for which we use the far-wing line shape  
 249 of Hartmann et al. (2002), adapted to an  $\text{H}_2$  atmosphere. In a second step,  
 250 we discretize these spectra in large bands (typically  $100\text{--}300\text{ cm}^{-1}$  wide) and  
 251 use the KDISTRIBUTION code (Eymet et al., 2016) to compute correlated- $k$   
 252 coefficients  $k(g)$  for each spectral band and each (T,p) value. We sample the  
 253 cumulative probability  $g$  with 8 Gauss integration points from 0 to 0.95 and  
 254 another 8 points from 0.95 to 1. The spectral discretization and the number  
 255 of bandwidths is a compromise between accuracy (which increases when small  
 256 spectral intervals are chosen) and computation time. After multiple tests, we  
 257 have selected 20 bands in the thermal infrared ( $10\text{--}3200\text{ cm}^{-1}$  or  $3\text{ }\mu\text{m} - 1$   
 258  $\text{mm}$ ) and 25 bands in the visible and near-infrared ( $2000\text{--}33000\text{ cm}^{-1}$  or  $0.3\text{--}$   
 259  $5\text{ }\mu\text{m}$ ). When running a radiative-convective simulation, these  $k$ -coefficients  
 260 are interpolated at each time step to the local temperature and to the pres-  
 261 sure grid of the radiative transfer model. All radiative-convective simulations  
 262 presented in this paper use a pressure grid consisting of 64 levels between 3  
 263 and  $10^{-6}$  bar.

### 264 2.3 Updates on methane spectroscopy

265 Spectroscopic line parameters are extracted from the HITRAN 2016 database  
 266 (Gordon et al., 2017). However, the  $\text{CH}_4$  linelist is known to be incomplete  
 267 beyond  $7,900\text{ cm}^{-1}$ ; in particular, a methane absorption band at  $1\text{ }\mu\text{m}$  is  
 268 missing entirely, which could lead to an underestimation of the atmospheric  
 269 heating rates. To fill this gap, we complete the HITRAN 2016 methane

270 linelist with a recent linelist based on *ab initio* calculations (Rey et al., 2016,  
 271 2018). This list is available in the 0–12,000  $\text{cm}^{-1}$  range and contains posi-  
 272 tion, energy and intensity for nearly 3.5 millions of transitions (assuming a  
 273 intensity cut-off of  $10^{-28}$   $\text{cm}/\text{molecule}$ ), where the HITRAN 2016 database  
 274 contains about 340,000 transitions. In order to limit the computation time,  
 275 and because the HITRAN 2016 methane database is thought to be reliable up  
 276 to 7,900  $\text{cm}^{-1}$ , we choose to combine the two linelists, using the spectroscopic  
 277 parameters of Rey et al. (2016) only beyond 7,900  $\text{cm}^{-1}$ . Furthermore, we  
 278 now include the transitions of the isotopologues  $^{13}\text{CH}_4$  and  $\text{CH}_3\text{D}$  that were  
 279 previously neglected by Guerlet et al. (2014). The isotopic ratio  $^{13}\text{C}/^{12}\text{C}$  is  
 280 set to the terrestrial value (0.011) in agreement with Galileo measurements  
 281 (Niemann et al., 1998), and the ratio  $\text{CH}_3\text{D}/\text{CH}_4$  to  $7.79 \times 10^{-5}$  (Lellouch  
 282 et al., 2001). The spectroscopic line data of  $^{13}\text{CH}_4$  and  $\text{CH}_3\text{D}$ , for the spec-  
 283 tral domain 0–12,000  $\text{cm}^{-1}$ , also come from *ab initio* calculations by Rey  
 284 et al. (2016), which are more exhaustive than HITRAN 2016.

285 Figure 1 shows the comparison between absorption coefficient spectra in  
 286 the visible range computed using the HITRAN 2016 database (considering  
 287  $^{12}\text{CH}_4$  only) with the new combination of the HITRAN 2016 and Rey et al.  
 288 (2018) linelists for  $^{12}\text{CH}_4$ ,  $^{13}\text{CH}_4$  and  $\text{CH}_3\text{D}$ . This figure illustrates the im-  
 289 portant addition of the  $^{12}\text{CH}_4$  Rey et al. (2018) linelist to HITRAN 2016  
 290 beyond 7,900  $\text{cm}^{-1}$ , as well as the contribution of  $\text{CH}_3\text{D}$  that features emis-  
 291 sion bands at 1,100  $\text{cm}^{-1}$  (not shown), 2,200  $\text{cm}^{-1}$  and 3,500  $\text{cm}^{-1}$ .  $^{13}\text{CH}_4$   
 292 lines are not visible in this figure as their main absorption bands are mingled



with  $^{12}\text{CH}_4$ . Regarding the impact on the equilibrium temperature profile, using the  $^{12}\text{CH}_4$  linelist from Rey et al. (2018) beyond  $7,900\text{ cm}^{-1}$  increases the heating rates by 10% to 20% compared to using HITRAN 2016 alone in the range  $0\text{--}12,000\text{ cm}^{-1}$ . This corresponds to a stratospheric warming between 2 and 3.5 K, the maximum lying near 10-20 mbar. The addition of the two methane isotopologues yields a modest increase of  $\sim 1\text{ K}$ .

We recall that for the Saturn model, Guerlet et al. (2014) used a combination of HITRAN 2012 (similar to HITRAN 2016 as far as  $\text{CH}_4$  is concerned) up to  $7,800\text{ cm}^{-1}$  with another set of  $k$ -distribution coefficients computed in the range  $7,800\text{--}25,000\text{ cm}^{-1}$  based on the Karkoschka and Tomasko (2010) methane band model. In the Guerlet et al. (2014) study, we concluded that the amount of heating by methane beyond  $7,800\text{ cm}^{-1}$  was significant, but we did not distinguish between the near infrared part ( $7,800\text{--}12,000\text{ cm}^{-1}$ ) and the visible part beyond  $12,000\text{ cm}^{-1}$ . In order to complete our study and evaluate the radiative heating resulting from absorption of visible solar photons, we computed a new set of  $k$ -distribution coefficients in the range  $12,000\text{--}25,000\text{ cm}^{-1}$  based on the Karkoschka and Tomasko (2010) data. We find that, as far as our Jupiter model is concerned, absorption by methane in this range has a negligible impact, warming the atmosphere by at most  $0.4\text{ K}$  at the 10-20 mbar level. This can be explained by the small volume mixing ratio of methane combined with its rapidly decreasing absorption coefficients beyond  $12,000\text{ cm}^{-1}$ .

Hence, in this paper we choose to work only with the HITRAN 2016 and

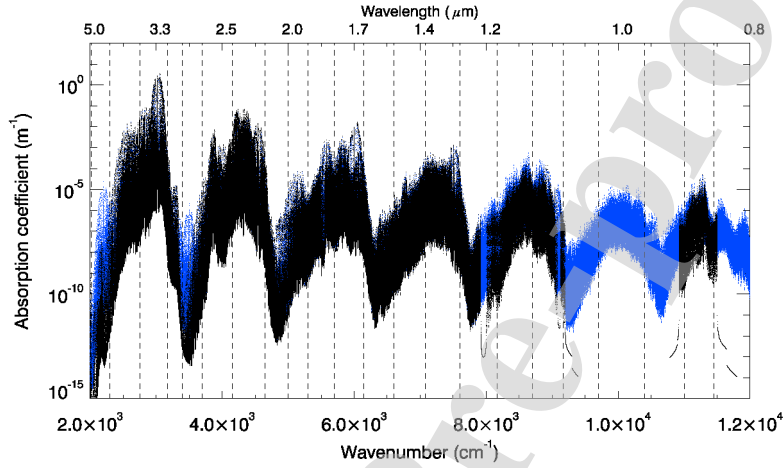


Figure 1: Absorption coefficient spectrum calculated for a pressure of 10 mbar and a temperature of 160K in the visible range from the HITRAN 2016 database (main  $^{12}\text{CH}_4$  isotope, in black), and also including Rey et al. (2018) linelists for  $^{12}\text{CH}_4$  (beyond  $7,900\text{ cm}^{-1}$ ),  $\text{CH}_3\text{D}$  and  $^{13}\text{CH}_4$  (in the range  $0\text{--}12,000\text{ cm}^{-1}$ ), in blue. The vertical dashed lines represent the limits of the 22 bands used for generating  $k$ -distribution coefficients.

316 Rey et al. (2018) linelists and neglect gaseous absorption in the visible part  
 317 (that is, beyond  $12,000\text{ cm}^{-1}$ ). The near infrared part of the spectrum is  
 318 discretized in 22 spectral intervals, shown in Figure 1, to which we add three  
 319 bands covering the visible part with zero gaseous opacity. These bands are  
 320 needed to contain cloud and aerosol opacity.

## 321 2.4 Gaseous abundances

322 In the present study, we neglect meridional variations of the trace species.  
 323 Instead, the  $k$ -tables are computed for a single volume mixing ratio vertical  
 324 profile for each species, reflecting Jupiter's average composition. We set the  
 325  $\text{H}_2$  volume mixing ratio to 0.863, the helium mixing ratio to 0.136 (Niemann  
 326 et al., 1998), and the mixing ratio of  $\text{CH}_4$  in the deep troposphere to the value  
 327 determined in situ by the Galileo probe mass spectrometer ( $2.07 \pm 0.5 \times 10^{-3}$ ,  
 328 Wong et al. (2004b)). We note that other values of the methane mixing ratio  
 329 have been reported from independent, remote sensing measurements, such  
 330 as Gautier et al. (1982). They determined a value of  $1.8 \pm 0.2 \times 10^{-3}$ , which  
 331 is consistent with the value of Wong et al. (2004b), within error bars.

332 The volume mixing ratio of  $\text{CH}_4$  decreases with altitude primarily due  
 333 to molecular diffusion in the upper stratosphere, and to a lesser extent due  
 334 to photo-dissociation by solar UV radiation (Gladstone et al., 1996; Moses  
 335 et al., 2005). The altitude level of the methane homopause on Jupiter is  
 336 estimated to lie in the range  $10^{-5}$  to  $10^{-6}$  bar based on stellar occultations  
 337 (Festou et al., 1981; Yelle et al., 1996; Greathouse et al., 2010). This is sig-  
 338 nificantly deeper than on Saturn, where the homopause level is estimated to  
 339 a few  $10^{-7}$  bar. This difference is explained by a much stronger eddy mixing  
 340 coefficient on Saturn compared to Jupiter (Moses et al., 2005). However,  
 341 the exact homopause level on Jupiter is not well constrained by observations  
 342 and differs among studies (Greathouse et al., 2010); it could also vary with  
 343 time and latitude. Similarly, uncertainties on the eddy mixing coefficient and

344 photodissociation rates map into a family of the methane vertical abundance  
 345 profile in Jupiter photochemical models (see for instance models A, B and C  
 346 of Moses et al. (2005)).

347 The choice of a vertical profile for methane (and other hydrocarbons) is  
 348 thus partly arbitrary and will influence the vertical energy deposition, hence  
 349 the resulting equilibrium temperature profile in the stratosphere, as was al-  
 350 ready reported by Zhang et al. (2013a) and Kuroda et al. (2014). We choose  
 351 to set the volume mixing ratio of the three hydrocarbons to the average of  
 352 the 1-D photochemical models A and C from Moses et al. (2005). This cor-  
 353 responds to an homopause level at  $\sim 1 \mu\text{bar}$ . Regarding  $\text{C}_2\text{H}_6$  and  $\text{C}_2\text{H}_2$ ,  
 354 we further scale these model profiles so that the hydrocarbon abundances at  
 355 1 mbar match the low to mid-latitude Cassini/CIRS observations of Nixon  
 356 et al. (2010):  $7.6 \times 10^{-6}$  for  $\text{C}_2\text{H}_6$ ,  $2.9 \times 10^{-7}$  for  $\text{C}_2\text{H}_2$ . For the purpose of  
 357 sensitivity tests, we also compiled a different set of  $k$ -tables with the hydro-  
 358 carbon profiles set to the photochemical model profiles used by Nixon et al.  
 359 (2007), which feature a deeper homopause level ( $\sim 10 \mu\text{bar}$ ). The different  
 360 hydrocarbon vertical profiles are illustrated in Fig. 2.

361 We present in Fig. 3 the impact of assuming different hydrocarbon pro-  
 362 files on the equilibrium temperature, based on aerosol-free 1-D radiative-  
 363 convection simulation for latitude  $20^\circ\text{N}$ . In the 1 mbar to  $10 \mu\text{bar}$  region, the  
 364 Nixon et al. (2007) photochemical model has  $\sim 1.5$  to 3 times more acetylene  
 365 than our combination of the Moses et al. (2005) models (but similar amounts  
 366 of ethane and methane), resulting in greater cooling rates and stratospheric

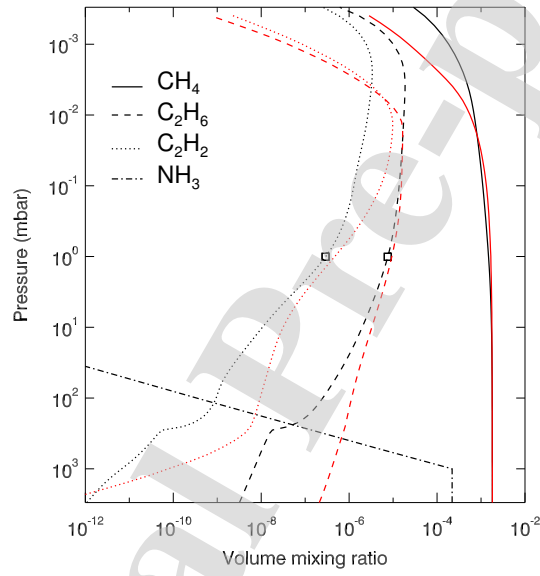


Figure 2: Vertical profiles for the volume mixing ratio of methane, ethane, acetylene and ammonia corresponding to an average of photochemical models “A” and “C” of Moses et al. (2005) (in black) or to the photochemical model used by Nixon et al. (2007) (in red). The  $C_2H_6$  and  $C_2H_2$  vertical profiles of Moses et al. are scaled to the abundances retrieved by Nixon et al. (2010) at 1 mbar and averaged between  $40^\circ S$  and  $40^\circ N$  (shown as squares).

temperatures 2 to 5K colder. Between 1 and 10  $\mu$ bar, the temperature calculated using the Nixon et al. (2007) hydrocarbons reaches a minimum, then increases with height. In this pressure range, all three hydrocarbons of the Nixon et al. (2007) model sharply decreases with altitude. This yields lower heating rates through lower absorption by  $\text{CH}_4$  (explaining the cold temperatures near 5  $\mu$ bar) but also lower cooling rates by hydrocarbon infrared emissions. As  $\text{C}_2\text{H}_6$  and  $\text{C}_2\text{H}_2$  decrease more sharply than  $\text{CH}_4$ , the net effect is a warming of the atmosphere between 5 and 1  $\mu$ bar.

We also evaluate the impact, on our equilibrium temperature profile, to a 30% decrease in both  $\text{C}_2\text{H}_6$  and  $\text{C}_2\text{H}_2$  mixing ratios with respect to our nominal hydrocarbon profiles based on the Moses et al. (2005) models. This 30% change reflects typical observed meridional and temporal variations at low to mid-latitudes (Melin et al., 2018). This yields a temperature increase of about 3K above the 10-mbar pressure level. This is in qualitative agreement with the work of Kuroda et al. (2014) who estimated a temperature change of  $\pm 8\text{K}$  when  $\text{C}_2\text{H}_6$  and  $\text{C}_2\text{H}_2$  were divided or multiplied by two.

Finally, we also quantify the impact of an increase of +50% in ethane mixing ratio, while acetylene is divided by two: this case study corresponds to what is observed at high latitudes compared to the equator (Nixon et al., 2010; Fletcher et al., 2016). In doing so, we evaluate the impact of neglecting actual meridional variations, in the (realistic) case where acetylene and ethane exhibit opposite trends. We find that the impact of increasing ethane while decreasing acetylene is rather small, as there is a partial compensa-

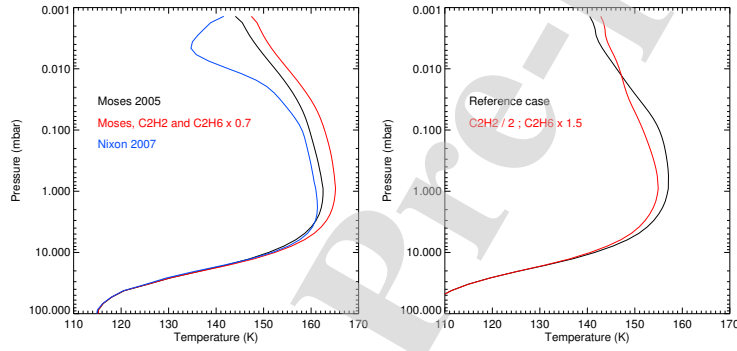


Figure 3: Equilibrium temperature profiles for different hydrocarbon mixing ratio profiles. Left: example at latitude  $20^{\circ}\text{N}$ ,  $L_s = 0$ , with the hydrocarbon abundances set to that of Nixon et al. (2007) (in blue) or to the average of model A and C of Moses et al. (2005) (in black) or the latter but with 30% less  $\text{C}_2\text{H}_2$  and  $\text{C}_2\text{H}_6$  (in red). Right: example at latitude  $60^{\circ}\text{N}$ ,  $L_s = 0$ , with the reference hydrocarbon abundances (the average of model A and C of Moses et al., 2005) (black line) or with a 50% increase in  $\text{C}_2\text{H}_6$  and a 50% decrease in  $\text{C}_2\text{H}_2$  (red line).

tion of the two competing effects (an increase in radiative cooling rates when ethane is increased, a decrease of it when acetylene is decreased). A similar conclusion was reached by Zhang et al. (2013a), as far as the Cassini/CIRS hydrocarbon retrievals were concerned. At 60°N and between 1 and 10 mbar, the resulting temperature profile (shown in Fig. 3) is 1 to 2 K colder than the nominal case, and is up to 4 K colder in the 1 to 0.05 mbar range. This is because ethane is a more efficient coolant than acetylene is in this pressure range. The two temperature profiles are then similar at and around 0.01 mbar. At even lower pressures, the temperature becomes slightly warmer than the nominal case. This is because acetylene cools more efficiently the upper stratosphere than ethane, as was already mentioned by Kuroda et al. (2014), so that a decrease by a factor of 2 of acetylene results in a net warming compared to the nominal case. Hence, we conclude that the impact of neglecting meridional variations in ethane and acetylene on stratospheric temperatures is of the order of 2–4 K.

In addition to hydrocarbons, we evaluate the influence of including ammonia ( $\text{NH}_3$ ). For the tropospheric temperatures encountered on Jupiter, ammonia is expected to condense at  $\sim 0.7$  bar ( $\sim 150\text{K}$ ) (Atreya et al., 1999). Following the vapour pressure curve, its mixing ratio rapidly decreases above this pressure level to become insignificant at tropopause levels. We set the ammonia "deep" mixing ratio (at 3 bar) to 250 ppm consistently with planet-average abundances measured by Juno at this pressure level (Li et al., 2017) and assume a fractional scale height of 0.15 above the 0.7 bar level (Nixon



et al., 2007). We find that including  $\text{NH}_3$  in our model yields a significant temperature increase of 10K in the troposphere. This temperature increase is caused by the absorption of near infrared solar light by  $\text{NH}_3$  and also by a small greenhouse effect. Indeed, adding ammonia increases the infrared opacity, especially beyond  $5\text{ }\mu\text{m}$ , as shown in Fig. 4. In consequence, thermal radiation emitted deep in the troposphere at long wavelengths is partly absorbed by ammonia in the mid-troposphere, which limits the cooling-to-space and warms the troposphere. We note that we have also tested including phosphine ( $\text{PH}_3$ ) with a deep abundance of  $6.0\times 10^{-7}$  and a fractional scale height of 0.3 (Nixon et al., 2007), but found a negligible impact on the thermal structure.

We choose here to keep the ammonia mole fraction constant with latitude. However, recent measurements made by the Juno microwave radiometer revealed highly variable ammonia concentrations, hinting at an ammonia-rich equatorial region (300–340 ppm in the 1–3 bar pressure range) and an ammonia-depleted region at  $10\text{--}20^\circ\text{N}$  (as low as 140 ppm at 1 bar) (Li et al., 2017). A sensitivity test where  $\text{NH}_3$  is decreased by 40% (150 ppm at the 1-bar level instead of 250 ppm) yields a small temperature decrease of 1K in the troposphere. Hence, the spatial variations derived from Juno should not significantly impact the thermal structure (in terms of direct radiative forcing). Having included ammonia in our Jupiter model does not challenge our previously published results on Saturn’s thermal structure (Guerlet et al., 2014). Indeed, the upper tropospheric volume mixing ratio of  $\text{NH}_3$  is 10 to

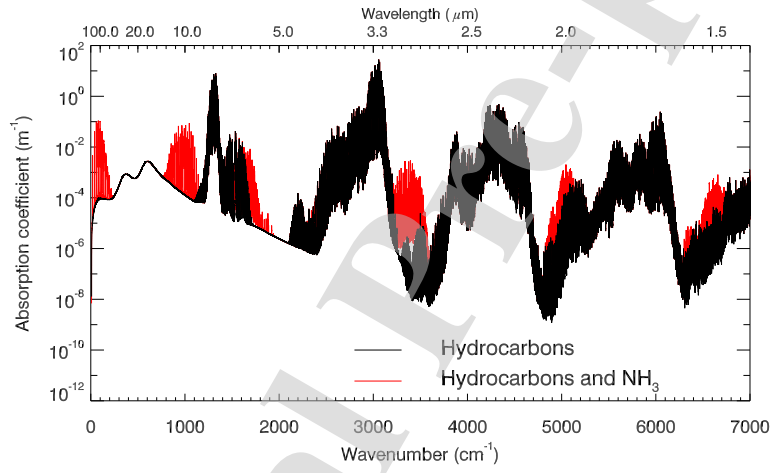


Figure 4: Absorption coefficient spectrum calculated for a pressure of 500 mbar and a temperature of 130K, including (in red) or not (in black) ammonia. Collision-induced absorption by  $\text{H}_2\text{-H}_2$  and  $\text{H}_2\text{-He}$  is included and is important in the 5–100  $\mu\text{m}$  range.

100 times lower on Saturn than on Jupiter, due to a deeper condensation level ( $\sim 1.4$  bar), and we find that including  $\text{NH}_3$  does not have a significant impact on Saturn's upper tropospheric temperatures.

## 2.5 Treatment of tropospheric clouds and aerosols

Cloud and haze particles are expected to play a key role in the radiative budget of Jupiter's troposphere. Through their vertical distribution, microphysical and optical properties, they control the local absorption of solar radiation at different depth, hence the temperature and heat redistribution. Many studies have attempted to characterize their physical and chemical properties from remote sensing measurements, with more or less agreement between them due to the complexity of such ill-posed inverse problems, with non-unique solutions. A complete review on the cloud and haze observational constraints would be beyond the scope of this paper; instead we summarize below the main findings of the cloud and haze radiative impact in the upper troposphere relevant to our study, at pressures less than 2–3 bar.

### 2.5.1 Observational constraints

There is an overall consensus that, in order to reproduce both visible and thermal infrared imaging data, a combination of a diffuse haze comprising small particles ( $0.3\text{--}2\ \mu\text{m}$ ) located above a compact cloud comprising larger particles ( $3\text{--}100\ \mu\text{m}$ ) is needed (*e.g.* West et al. (1986) from Pioneer data, Irwin et al. (2001) using Galileo/NIMS spectra, Wong et al. (2004a) using

457 Cassini/CIRS data, Sromovsky and Fry (2018) using New Horizon/LEISA).

458 The location of the cloud deck is estimated to lie in the range 0.5–1.2 bar  
 459 depending on the studies, while the upper tropospheric haze likely extends  
 460 up to 150–300 mbar, *i.e.* near the tropopause. Thermochemical equilibrium  
 461 models predict that ammonia condenses at  $\sim 700$  mbar, while ammonium  
 462 hydrosulfide ( $\text{NH}_4\text{SH}$ ) is expected to form another cloud layer at  $\sim 2$  bar  
 463 (Atreya et al., 1999). However, the spectroscopic signatures of  $\text{NH}_3$  ice at  
 464  $2\text{ }\mu\text{m}$ ,  $9.4\text{ }\mu\text{m}$  and  $26\text{ }\mu\text{m}$  have been rarely observed, and Baines et al. (2002)  
 465 showed that spectrally identifiable ammonia clouds cover less than 1% of  
 466 Jupiter’s globe. Rather, Sromovsky and Fry (2010) suggest that the haze  
 467 layer consists of small ammonia-coated particles overlying a cloud layer of  
 468  $\text{NH}_4\text{SH}$  ice particles at  $\sim 600$  mbar, or that several layers of  $\text{NH}_3$  and  $\text{NH}_4\text{SH}$   
 469 ice particles coexist, which would explain the lack of strong  $\text{NH}_3$  absorption  
 470 features in the infrared. A similar conclusion was reached by Giles et al.  
 471 (2015), who used Cassini/VIMS data between  $4.5$  and  $5.2\text{ }\mu\text{m}$  to constrain  
 472 Jupiter’s cloud structure. The authors find that VIMS observations can be  
 473 modeled using a compact, highly reflecting cloud layer located at a pressure  
 474 of 1.2 bar or lower, with spectrally flat optical properties in this spectral  
 475 range. Indeed, setting the refractive index to that of pure  $\text{NH}_3$  or  $\text{NH}_4\text{SH}$  ice  
 476 particles could not fit VIMS observations, for any particle sizes in the range  
 477  $1\text{--}40\text{ }\mu\text{m}$ .

478 A few observational constraints exist on haze and cloud particles optical  
 479 properties: Pioneer observations analyzed by Tomasko et al. (1978) require

highly reflecting particles at 0.44 and 0.6  $\mu\text{m}$ , with single scattering albedo higher than 0.95 for the haze, and higher than 0.98 for the cloud particles. Typical cumulative optical depths measured in the visible (0.75  $\mu\text{m}$ ) vary from 1 to 5 above the 500-mbar level (that can reasonably be attributed to the haze opacity), and vary between 5 and 20 above the 1-bar level (see Sromovsky and Fry, 2010, and references therein). In the near infrared, haze optical depths varying between 0.5 and 5 have been derived at 2  $\mu\text{m}$  (Irwin et al., 2001; Kedziora-Chudczer and Bailey, 2011). The optical depth variations between cloudy zones and less opaque belts likely stem from cloud optical depth variations (found to lie between 8 and 22 at 5  $\mu\text{m}$ , Giles et al., 2015) rather than variations of the haze itself.

### 2.5.2 Cloud model and sensitivity studies

Our goal here is to set up an effective cloud and haze model that would reproduce Jupiter's albedo, thermal structure and be consistent with the observed visible and infrared cloud optical depths and physical properties at a global scale. We emphasize that this effective model is not meant for comparisons to detailed spectroscopic observations, but rather is meant to account for the radiative forcing of cloud and haze particles and their role in the radiative budget. In what follows, we assume a two-layer cloud structure with an extended, upper haze located above a compact cloud and test the sensitivity to varying the cloud and haze composition (optical constants), optical depth, particle sizes and the altitude of the cloud deck. We assume spherical

particles and compute the optical properties (single scattering albedo, extinction coefficient and asymmetry factor) with a Mie scattering code. Four compositions are tested:

1. pure  $\text{NH}_3$  ice particles, with optical constants from Martonchik et al. (1984);
2. pure  $\text{NH}_4\text{SH}$  ice particles, with optical constants from Howett et al. (2007);
3. same material as our Saturn haze model (Guerlet et al. (2014), based on observational constraints from Karkoschka and Tomasko (1993) on Saturn);
4. particles with nearly grey optical constants.

Composition 4 has real and imaginary indices set arbitrarily close to that of  $\text{NH}_3$  except for smoother spectral features (since the sharp absorption features of  $\text{NH}_3$  are not observed) – reaching spectrally-flat in the visible range. This makes the imaginary index of composition 4 intermediate between compositions 1 and 3 for haze particles. The refractive indexes for the four kind of haze particles are compared in Figure 5. We note that  $\text{NH}_4\text{SH}$  particles (composition 2) are expected to be brighter than the other kind of particles, as a result of the low real index of  $\text{NH}_4\text{SH}$ , while the “Saturn” particles (composition 3) should be the most absorbing ones in the visible, owing to their higher imaginary index shortward of  $1\ \mu\text{m}$ .

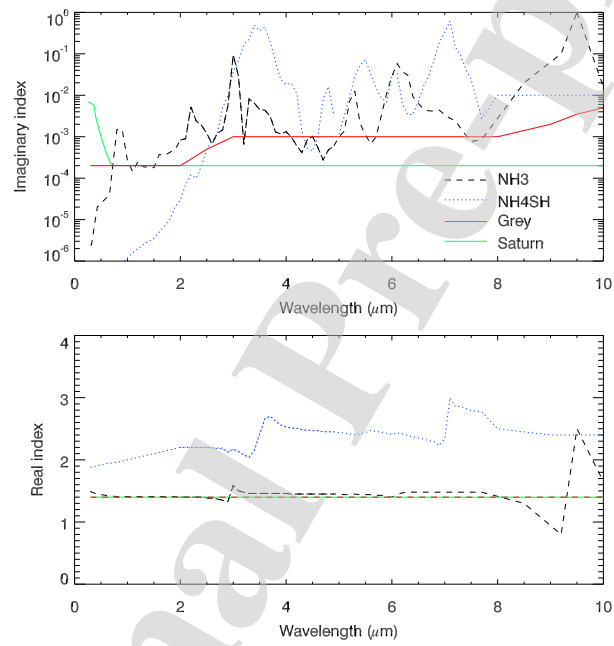


Figure 5: Imaginary (top) and real (bottom) refractive indexes of four assumed cloud or haze compositions. The “Saturn” real refractive index is the same as for the “grey” particle type.

523 In order to study the impact of the cloud properties on the planetary  
 524 albedo, we first perform 1-D radiative-convective simulations for globally-  
 525 averaged conditions. The planetary albedo is defined as  $1 - \frac{ASR}{ISR}$ , where ISR  
 526 stands for incoming solar radiation and ASR for absorbed solar radiation,  
 527 both quantities being evaluated globally. This value of modeled planetary  
 528 albedo is to be compared to the observed Bond albedo of 0.50 according to Li  
 529 et al. (2018b). In this first set of simulations, the aerosol vertical structure is  
 530 fixed with a reference cloud deck at 840 mbar and a scale height of 0.2 times  
 531 the atmospheric scale height (to be consistent with previous observations of  
 532 a compact cloud) along with an upper haze extending from 660 to 150 mbar,  
 533 with a scale height set to the atmospheric scale height. Table 1 presents a  
 534 set of results for different couples of cloud and haze composition, varying the  
 535 haze particle size between 0.5 and 2  $\mu\text{m}$ , the haze integrated optical depth  
 536 at 0.75  $\mu\text{m}$  between 1 and 4, the cloud particle size between 10 and 20  $\mu\text{m}$ ,  
 537 and the cloud integrated optical depth in the visible between 7 and 15 (only  
 538 22 out of the 108 combinations tested are shown in Table 1).

539 Overall, we find that all cases considering a pure  $\text{NH}_4\text{SH}$  cloud lead to  
 540 a too bright albedo ( $>0.55$ ), regardless of the assumptions on cloud opti-  
 541 cal depth, cloud particle size or haze properties. Similarly, all cases with  
 542 “Saturn”-like haze particles combined with ammonia cloud particles result  
 543 in too dark albedos ( $\sim 0.4$ ), which is consistent with the high imaginary in-  
 544 dex of these haze particles in the visible. A combination of a dark “Saturn”  
 545 haze with a bright  $\text{NH}_4\text{SH}$  cloud leading to a  $\sim 0.5$  albedo might be found,



546 but we choose to discard solutions with the “Saturn” haze as several studies  
 547 (e.g., Tomasko et al., 1978) suggest that Jupiter’s haze particles must have  
 548 a larger single scattering albedo than Saturn’s. The “grey” and  $\text{NH}_3$  haze  
 549 particles considered here are in better agreement with estimates of the single  
 550 scattering albedo by Tomasko et al. (1978). The latter study also constrained  
 551 the phase function of upper tropospheric cloud particles, which were found  
 552 to have a strong forward scattering lobe in the visible. The asymmetry param-  
 553 eter (computed from a Mie scattering code) of our cloud particles lies in the  
 554 range 0.8 to 0.85, which also indicate strong forward scattering.

555 Albedos comparable to that reported by Li et al. (2018b) are obtained  
 556 for combinations of “grey” and/or  $\text{NH}_3$  particles for the haze and cloud ma-  
 557 terial, with the condition that the haze optical depth amounts to 3–4. Dif-  
 558 ferent combinations of the nature of the haze and cloud particles ( $\text{NH}_3$ – $\text{NH}_3$ ,  
 559 grey–grey, grey– $\text{NH}_3$ ) give similar results, which is not surprising given their  
 560 similar optical constants. Hence, even though spectroscopic studies have  
 561 ruled out pure  $\text{NH}_3$  ice particles for the cloud composition, it seems that  
 562 assuming a  $\text{NH}_3$  or “grey” cloud does not impact greatly the overall energy  
 563 budget, and our model results are not very sensitive to one or the other type  
 564 of composition. We confirm that small haze particles ( $\sim 0.5 \mu\text{m}$ ) are needed  
 565 in order to reproduce the 3 to 4 times larger haze optical depth observed in  
 566 the visible compared to the near infrared: with the haze optical depth set  
 567 to 4 at  $0.75 \mu\text{m}$  for ammonia or grey particles, the optical depth amounts to  
 568  $\sim 1$  at a wavelength of  $2 \mu\text{m}$ , which is compatible with observations by Irwin

et al. (2001).

The heat deposition differs depending on the cloud and haze composition, integrated optical depth and altitude of the cloud deck, as illustrated in Fig 6. For scenarios with bright  $\text{NH}_4\text{SH}$  clouds, the heating rate decreases moderately within the haze and cloud layer, while for scenarios with ammonia or “grey” cloud particles, the heat deposition reaches a local maximum within the cloud layer. Actually, because the cloud optical depth is large, the maximum heat deposition occurs above the cloud deck. In other words, at the cloud deck, there is little visible radiation left to be absorbed. Fig 6 also illustrates the larger heating rate resulting from the absorption by “Saturn”-like haze particles compared to “grey” particles. We also note that excluding completely haze and cloud opacity in the model results in unrealistic albedo (0.15) and heat deposition profile.

We then evaluate the impact of changing the cloud optical depth as well as the altitude of the cloud deck on the temperature. In our “grey haze,  $\text{NH}_3$  cloud” scenario, increasing the cloud optical depth from 7 to 15 results in a 3 K warming of the troposphere (below the 300-mbar level), as absorption of visible solar photons increases. As stated in the introduction, we note that the opposite trend is actually observed on Jupiter: the cloudy equatorial zone is found to be 2 to 5 K colder than warmer, less cloudy, equatorial belts (e.g., Fletcher et al., 2016)). This reinforces the idea that zones are regions of upwelling (see for instance Gierasch et al., 1986) where adiabatic cooling dominates over radiative heating. Finally, moving the cloud deck from 840

Table 1: Cloud and haze properties along with the planetary albedo computed from globally-averaged 1-D radiative-convective simulations. Bold figures highlight our favored scenario, for which the albedo is close to 0.5, as reported by Li et al. (2018b).

Cloud			Haze			Bond Albedo
type	size (in $\mu\text{m}$ )	$\tau$ cloud at 750 nm	type	size (in $\mu\text{m}$ )	$\tau$ haze at 750 nm	
NH <sub>4</sub> SH	15.00	7.00	Grey	<b>0.50</b>	2.00	0.59
NH <sub>4</sub> SH	15.00	7.00	Grey	1.00	2.00	0.59
NH <sub>4</sub> SH	15.00	7.00	Grey	0.50	4.00	0.61
NH <sub>4</sub> SH	15.00	10.00	Grey	0.50	2.00	0.63
NH <sub>4</sub> SH	15.00	7.00	NH <sub>3</sub>	1.00	2.00	0.59
NH <sub>4</sub> SH	15.00	7.00	“Saturn”	1.00	2.00	0.55
NH <sub>4</sub> SH	15.00	15.00	“Saturn”	1.00	2.00	0.62
NH <sub>3</sub>	15.00	7.00	“Saturn”	1.00	2.00	0.39
NH <sub>3</sub>	15.00	7.00	“Saturn”	1.00	4.00	0.43
NH <sub>3</sub>	15.00	15.00	“Saturn”	1.00	2.00	0.40
NH <sub>3</sub>	15.00	15.00	“Saturn”	1.00	4.00	0.43
NH <sub>3</sub>	15.00	4.00	Grey	0.50	2.00	0.41
NH <sub>3</sub>	15.00	10.00	Grey	0.50	2.00	0.43
NH <sub>3</sub>	15.00	7.00	Grey	1.00	2.00	0.42
NH <sub>3</sub>	15.00	7.00	Grey	2.00	2.00	0.42
NH <sub>3</sub>	10.00	7.00	Grey	0.50	2.00	0.44
NH <sub>3</sub>	20.00	7.00	Grey	0.50	2.00	0.41
NH <sub>3</sub>	15.00	7.00	Grey	0.50	4.00	<b>0.48</b>
NH <sub>3</sub>	10.00	10.00	Grey	0.50	4.00	<b>0.50</b>
NH <sub>3</sub>	10.00	15.00	Grey	0.50	4.00	<b>0.51</b>
NH <sub>3</sub>	10.00	10.00	NH <sub>3</sub>	0.50	4.00	<b>0.51</b>
Grey	10.00	10.00	Grey	0.50	4.00	<b>0.48</b>

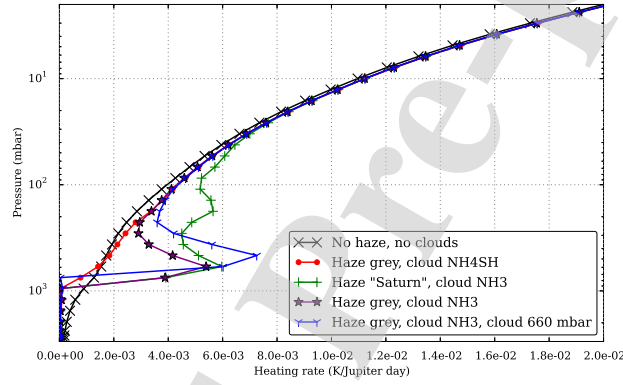


Figure 6: Vertical profiles of heating rates (in Kelvin per Jupiter day) due to absorption of solar radiation in the visible and near infrared for different cloud and haze particles, for globally-averaged conditions and  $L_s=180^\circ$ . In this example, the haze and cloud optical depth at  $0.75 \mu\text{m}$  are set to 4 and 10, respectively, and the haze and cloud particle sizes to  $0.5$  and  $10 \mu\text{m}$ . The cloud deck is set at 840 mbar except for one case with a slightly shallower cloud deck, in blue.

to 660 mbar results in warming the troposphere by 5 K (while having little impact on the albedo, of the order of a 2% change), as more solar light is absorbed at higher altitudes. Because the resulting temperature at the 1-bar level is closer to observations (detailed in section 3) when setting the cloud deck at 840 mbar, we choose to keep this setting throughout this paper.

In this section, we have documented the impact of different cloud and haze scenario on the tropospheric temperature and albedo. One has to keep in mind that modifications of the cloud or haze optical depth and altitude distribution will influence the tropospheric temperature by a few kelvins, just like modifications of the ammonia and hydrocarbon mixing ratio will also influence the temperature. Setting a realistic meridional profile of these variables is beyond the scope of the current project: not only observational constraints are limited, but feedbacks with meridional circulation – not yet included – are expected to play an important role as well. Hence, in the goal of setting an effective parametrization, able to reproduce the mean tropospheric temperature and global albedo, our nominal scenario is the following: a haze layer with an integrated optical depth of 4 in the range 660–150 mbar, “grey” particles of radius  $0.5\ \mu\text{m}$  on top of a  $\text{NH}_3$  cloud (or indifferently a “grey” cloud) with  $10\text{-}\mu\text{m}$  particles, a cloud deck at 840 mbar with an integrated optical depth at 750 nm of 15.

## 612 2.6 Stratospheric aerosols

### 613 2.6.1 Observational constraints and motivation

614 In addition to the tropospheric cloud and aerosol layers described above, we  
615 take into account two stratospheric haze layers:

- 616 1. We include an optically thin stratospheric layer comprising small spher-  
617 ical particles ( $0.2\text{--}0.5\ \mu\text{m}$ ) with an integrated optical depth set to 0.02  
618 in the NIR and UV, as constrained by Zhang et al. (2013b). Their re-  
619 fractive index have been constrained in the same study, with imaginary  
620 parts at 255 nm and 900 nm estimated to 0.02 and 0.001, respectively.  
621 This haze is uniform with latitude and extends from the tropopause to  
622 the upper stratosphere, with a scale height equal to the atmospheric  
623 scale height. Its impact on the stratospheric temperature is  $<0.5\text{K}$ .
- 624 2. We include another layer that is not uniform with latitude and is more  
625 absorbant in particular in the UV, described further below.

626 The addition of this second kind of aerosol is motivated by the observations  
627 of dark polar hoods at near-UV wavelength (Hord et al., 1979; Tomasko  
628 et al., 1986), which have been attributed to a stratospheric haze layer. This  
629 haze is both forward scattering and strongly polarizing, which implies that  
630 it is composed of aggregate particles similar to Titan's haze particles (West  
631 and Smith, 1991). The favored scenario for their formation is through pre-  
632 cipitation of energetic particles in Jupiter's upper atmosphere in its auroral

regions (Pryor and Hord, 1991), thought to be responsible for the production of heavy hydrocarbons (Wong et al., 2003). According to chemical and microphysical models, these hydrocarbons can condense and form fractal aggregates through coagulation processes (Friedson et al., 2002).

Recently, Zhang et al. (2013b) brought new constraints on the size, shape, vertical and meridional distribution of these stratospheric aerosols by combining ground-based near-IR spectra from Banfield et al. (1998) and multiple phase angle images from the Cassini Imaging Science Subsystem (ISS). The authors first derive the vertical profile of the aerosol mixing ratio at different latitudes and find that the stratospheric haze layer resides at a pressure of 50 mbar at low latitudes and  $\sim 20$  mbar at high latitudes ( $60\text{--}70^\circ$ ). Regarding the aerosol sizes and shapes, ISS observations can be fitted with small sub-micron spherical particles at low latitudes (which corresponds to the first type of stratospheric haze layer included in our model). Poleward of  $30^\circ\text{N}$  and  $45^\circ\text{S}$ , ISS observations are consistent with fractal aggregates with a fractal dimension of 2, corresponding to an effective radius of about  $0.7\mu\text{m}$ . Zhang et al. (2013b) also constrain the real and imaginary part of the refractive index of the fractal aggregates at two wavelengths in the UV and near-IR (at 255 nm and 900 nm). They derive a family of solutions, with different plausible combinations of refractive indexes, number and radius of monomers. For instance, their reference case corresponds to an imaginary index  $n_i$  of 0.02 at 255 nm and  $10^{-3}$  at 900 nm and aggregates comprising 1000 monomers with a 10-nm radius. Other solutions can match the Cassini/ISS

656 observations. Two extreme cases are : aggregates comprising 100 monomers  
 657 of 40 nm, with higher optical refractive indexes (0.08 in the UV and  $5 \times 10^{-3}$   
 658 in the NIR); or aggregates comprising 10000 monomers of 5 nm, with lower  
 659 optical indexes ( $6.10^{-3}$  in the UV and  $2.10^{-4}$  in the NIR).

660 Based on these observational constraints, Zhang et al. (2015) show that  
 661 this haze dominates the radiative heating budget at middle and high latitudes  
 662 in Jupiter's middle stratosphere, with a contribution of the haze reaching up  
 663 to 10 times the heating rate due to  $\text{CH}_4$  alone in the 10–20 mbar pressure  
 664 range. This haze can also cool the atmosphere through its infrared emission  
 665 (see also Guerlet et al. (2015) for a Saturn counterpart). Hence, radiative  
 666 heating and cooling by the polar haze appears to be a key component to  
 667 be included in any radiative-convective equilibrium model of Jupiter. The  
 668 parametrization of this haze in our model is detailed below.

## 669 2.6.2 Parametrization of the aerosol properties

670 The optical properties (extinction coefficient, scattering albedo and asym-  
 671 metry factor) of fractal aggregates haze particles are computed using a semi-  
 672 empirical model from Botet et al. (1997). This model employs the mean-field  
 673 approximation in the case of scattering of an electromagnetic wave by a clus-  
 674 ter of monosized spheres. We compute the optical properties for aggregates  
 675 with a fractal dimension of 2, and for the three aforementioned scenario deter-  
 676 mined by Zhang et al. (2013b) regarding the number and radius of monomers.  
 677 From UV to NIR, the real index is set to 1.65, similar to the mean value of



678 Zhang et al. (2015). The imaginary refractive indexes were set to the three  
 679 set of values determined by Zhang et al. (2013b) at 255 and 900 nm – one for  
 680 each combination of number and radius of monomers – with a logarithmic  
 681 interpolation in between as in Zhang et al. (2015). In the thermal infrared,  
 682 due to the lack of observational constraints, we adopt the real and imaginary  
 683 index of Vinatier et al. (2012) derived from Cassini/CIRS observations of  
 684 Titan’s stratospheric haze, which present striking similarities with Saturn’s  
 685 polar haze (Guerlet et al., 2015). For this given set of refractive index, op-  
 686 tical properties in the thermal infrared are computed three times, for the  
 687 three aforementioned values of number and radius of monomers. This en-  
 688 sures consistency in the considered scenario. Finally, we also include a case  
 689 where tholins-like properties are assumed (Khare et al., 1984). They are not  
 690 considered a very relevant analog for Jupiter’s haze – they do not match the  
 691 observed properties of Titan’s haze (Vinatier et al., 2012), let alone Saturn’s  
 692 haze – but this test is useful for sensitivity studies. For this test, we only  
 693 consider the scenario with 1000 monomers of 10-nm radius and we actually  
 694 divide by 2 the refractive index derived from the laboratory experiments of  
 695 Khare et al. (1984) to better match observations by Vinatier et al. (2012).  
 696 These different refractive index are summarized in figure 7 and our four sce-  
 697 narios summarized in table 2. We will discuss the impact of these different  
 698 properties on the thermal structure in the next section.

699 Regarding the meridional variations of the polar haze optical thickness,  
 700 we build a meridional profile based on that retrieved by Zhang et al. (2013b),

Table 2: Description of the 4 sets of parameters used to generate the stratospheric polar haze optical properties.  $N_{mono}$  is the number of monomers in the aggregates,  $r_{mono}$  the radius of the monomers, Im stands for the imaginary index.

Scenario	$N_{mono}$	$r_{mono}$	Im(255 nm)	Im(900 nm)	Im(Thermal infrared)
1	10000	5 nm	$6.10^{-3}$	$2.10^{-4}$	Vinatier et al. (2012)
2	1000	10 nm	$2.10^{-2}$	$1.10^{-3}$	Vinatier et al. (2012)
3	100	40 nm	$8.10^{-2}$	$5.10^{-3}$	Vinatier et al. (2012)
4	1000	10 nm	$2.10^{-2}$	$1.10^{-3}$	Khare et al. (1984)

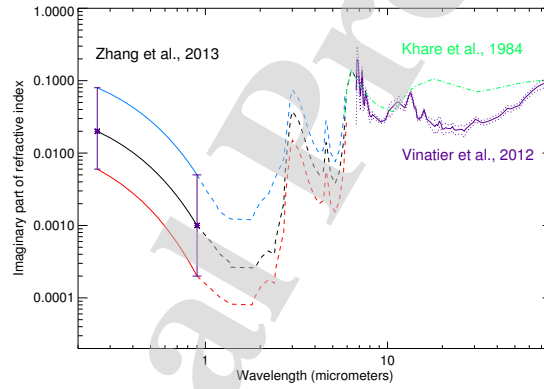


Figure 7: This figure illustrates the imaginary indexes used for generating optical properties of the stratospheric polar haze. Four sets of optical properties were computed based on different imaginary index, radius and number of monomers. We adopt the values constrained by Zhang et al. (2013b) at NIR and UV wavelength, and Vinatier et al. (2012) (in purple) or Khare et al. (1984) (divided by 2 here, green line) in the thermal infrared. Between 1 and 7  $\mu\text{m}$ , due to the lack of observing constraints, these values are interpolated, following the wavelength-dependence derived from tholins experiments by Khare et al. (1984).

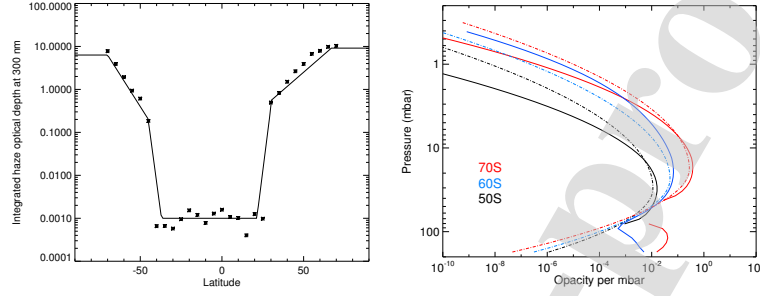


Figure 8: Left: Meridional variation of the stratospheric haze optical thickness at 300 nm integrated above the 80 mbar level as derived from Zhang et al. (2013b) (stars) compared to the values adopted in our model (dashed line). Right: Opacity vertical profiles at 300 nm derived from Zhang et al. (2013b) retrieved number density profiles, shown at three latitudes (solid lines), compared with those adopted in our model (dashed lines).

with a slightly smoother transition at mid-latitudes where the optical depth varies by several orders of magnitude over a few degrees of latitude. The integrated haze optical thickness as a function of latitude adopted in our model is compared with Zhang et al. (2013b) retrievals in Figure 8. Poleward of 70° (the highest latitude observed in Zhang et al., 2013b), we simply assume that the haze optical depth is equal to that at 70°.

We choose to parameterize the aerosol opacity vertical profile with a skewed gaussian profile:

$$\tau(p) \propto \exp(-(H \times \ln(p/p_m))^2 / 2\delta_1^2) / (\delta_1 + \delta_2) \quad (1)$$

with  $\tau$  the opacity per mbar,  $p$  the pressure,  $p_m$  parametrizing the pressure

level where the optical depth is maximum and  $\delta_1$  and  $\delta_2$  parametrizing the  
 width and skewness of the profile. This function reproduces well, to first  
 order, the vertical profile of the haze opacity derived from number density  
 profiles retrieved by Zhang et al. (2013b) (see Fig. 8). We vary  $p_m$  linearly  
 with latitude to capture the fact that the haze layer shifts from  $\sim 40$  mbar  
 to 20 mbar between  $\sim 45$  and  $70^\circ$ , as derived by Zhang et al. (2013b). In  
 our model, the opacity profile is then normalized at each latitude bin so that  
 the optical depth integrated above the 80 mbar level matches the merid-  
 ional profile in Fig. 8. The resulting vertical profiles of the haze opacity are  
 shown at three latitudes in Figure 8, along with those retrieved by Zhang  
 et al. (2013b). We note that Zhang et al. (2013b) retrievals suggest that,  
 poleward of  $60^\circ$ , the tropospheric aerosol layer shifts to lower pressure levels  
 ( $\sim 100$  mbar instead of 200 mbar), which we did not take into account (our  
 tropospheric layer extends up to 180 mbar at all latitudes).

### 2.6.3 Radiative impact of the haze

As reported by Zhang et al. (2015), we confirm that including the polar haze  
 results in strongly enhanced heating rates in the middle stratosphere, mostly  
 between  $40$  and  $70^\circ\text{N}$  and  $50$  and  $70^\circ\text{S}$ . Figure 9 shows an example at  $60^\circ\text{S}$ ,  
 $L_s = 0^\circ$ , where the heating rate is increased by a factor of four to six at  
 the 10-mbar pressure level when stratospheric aerosols are included, which  
 is in overall agreement with the 5 to 10 times enhancement factor reported  
 by Zhang et al. (2015). Using our radiative-convective equilibrium model,

we can go further and evaluate for the first time the impact of the polar  
 haze on the equilibrium temperature profile. We find that the temperature  
 is very sensitive to the polar haze properties. At this latitude and season,  
 considering scenario 2 or 3, the impact of the polar haze is to significantly  
 warm the stratosphere by 20K to 30K in the 10 to 30-mbar pressure range  
 (see Figure 9). This effect decreases with altitude, amounting to 3-5K at  
 the 1-mbar pressure level. If scenario 4 is considered (same as scenario 2 but  
 with the imaginary index of Khare et al. (1984) divided by 2, more absorbant  
 in the thermal infrared than that of Vinatier et al. (2012)), the polar haze  
 net effect is a moderate warming (10K) that is maximum near the 30-mbar  
 pressure level. Finally, scenario 1 results in temperatures changes of the  
 order of +5K near the 30-mbar level and -5K near the 5-mbar level. Indeed,  
 in this case, a net cooling of the middle stratosphere is obtained despite the  
 increase in heating rates. This is explained by the simultaneous increase  
 in cooling rates due to the polar haze. We also note that despite similar  
 heating rates in scenario 2 and 3, the equilibrium temperature profiles differ  
 significantly. There again, these differences result from different cooling rates:  
 even though the thermal infrared imaginary indexes of Vinatier et al. (2012)  
 are used in both scenario 2 and 3, the number and radius of monomers was  
 varied among these scenario. A smaller number of larger monomers (scenario  
 3) is more efficient in cooling the atmosphere than a larger number of smaller  
 monomers (scenario 2). Hence, it appears crucial to better characterize the  
 haze properties (their refractive index but also size and dimensions of the

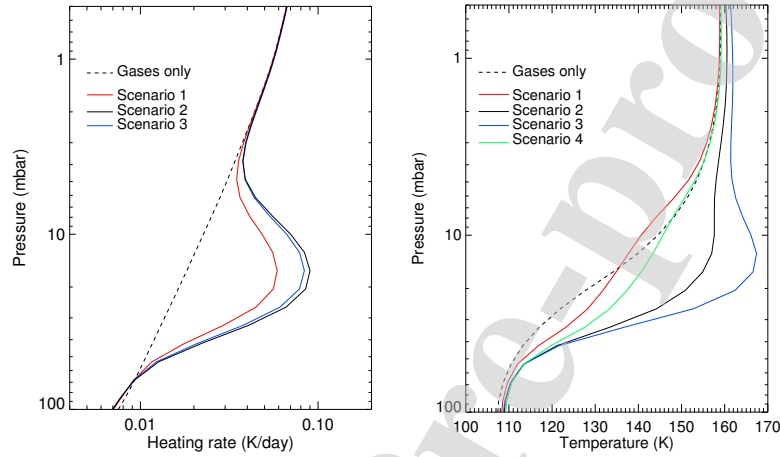


Figure 9: Daily-averaged heating rate, in Kelvin per Earth day (left) and temperature (right) vertical profiles at latitude  $60^{\circ}\text{S}$  and  $L_s = 0^{\circ}$ . Dashed black lines correspond to a case without the polar haze contribution, colored lines refer to different polar haze scenarios described in table 2. Scenario 4 is not shown in the heating rate figure as it is similar to scenario 2.

aggregates) in order to realistically model their radiative impact.

At high latitudes ( $> 75^{\circ}$ ) the net radiative impact of the polar haze depends on season: during winter, the net effect of scenario 2 is to cool the atmosphere, by typically 10K between 20 and 2 mbar. This is easily explained by the fact that the solar insolation is near zero at this season and location, while the aerosol layer still emits longwave radiation. On the other end, over the summer pole, the net effect can be an important warming (10–15K) between 20 and 2 mbar. Comparisons with observations, using simulations including or not the polar haze, are shown in the next section.

### 764 3 Thermal structure and comparisons to ob- 765 servations

#### 766 3.1 Internal heat flux and tropospheric temperature

767 Before presenting in detail the results of our radiative-convective model at  
768 equilibrium, we address the issue of the tropospheric equator-to-pole temper-  
769 ature gradient. Indeed, it is well known since the Pioneer 10 and 11 era that  
770 Jupiter exhibits no significant latitudinal gradient of temperature or emitted  
771 thermal infrared flux at the 1-bar level (Ingersoll, 1976). However, assum-  
772 ing a uniform internal heat flux  $F_{cst} = 7.48 \text{ W.m}^{-2}$ , our radiative-convective  
773 model produces a strong temperature contrast of 28 K at 1 bar between the  
774 warmer equator (178 K) and colder poles (150 K) (see Figure 13). This is  
775 expected from such a radiative model, as solar insolation is maximum at low  
776 latitudes all year round (given Jupiter's low obliquity).

777 To explain the observed near-uniform tropospheric temperatures, several  
778 theories have been proposed. For instance, using a turbulent, 3-D deep con-  
779 vection model, Aurnou et al. (2008) finds that convective heat transfer by  
780 quasi-geostrophic thermal plumes results in an outward heat flow 2.5 to 3.2  
781 times greater at the poles than at equator. This latitudinal trend is consis-  
782 tent with the work of Pirraglia (1984) who tried to estimate the meridional  
783 variations of internal heat flux needed to reconcile the observed solar en-  
784 ergy deposition with the outgoing thermal radiation. On a different note,

785 with their General Circulation Model for Jupiter’s troposphere, Young et al.  
 786 (2019) found that even when considering a uniform internal heat flux, atmo-  
 787 spheric dynamics acts to balance the latitudinal-varying solar forcing. As a  
 788 consequence, the 1-bar equator-to-pole temperature gradient is reduced from  
 789 35 K with a radiative-convective version of their GCM to only 5 K when  
 790 using their full GCM with resolved atmospheric dynamics.

791 In order to emulate these effects in our radiative-convective model, we  
 792 test different functions to vary the internal heat flux  $F_{int}$  with latitude  $\theta$ , for  
 793 instance:

$$F_{int}(\theta) = 0.67 \times F_{cst} + 0.66 \times F_{cst} \times \sin^2(\theta) \quad (2)$$

$$F_{int}(\theta) = 0.5 \times F_{cst} + F_{cst} \times \sin^2 \theta \quad (3)$$

794 This *ad hoc* parametrization ensures a planet-average internal heat flux equal  
 795 to  $F_{cst}$  while setting an internal heat flux twice larger (eq. 2) or three times  
 796 larger (eq. 3) at the poles than at the equator. When using eq. 2, the equato-  
 797 rial temperature at the 1-bar level is now 9 K warmer than the poles (instead  
 798 of 28 K when a uniform internal heat flux is assumed). The associated out-  
 799 going thermal emission is only 8% larger at the equator than at 60° latitude,  
 800 which is consistent with Pirraglia (1984) observations, which extended to 60°  
 801 only. However, when using eq. 3, the temperature is actually 2 K warmer  
 802 (and the outgoing thermal emission 8% larger) at 60° than at the equator.  
 803 Hence, in what follows, we discuss the thermal structure obtained with eq. 2  
 804 which yields more realistic results. It is worth emphasizing here that the  
 805



temperature field at pressures lower than  $\sim 50$  mbar is not impacted by the hypothesis of a uniform or varying internal heat flux.

### 3.2 Thermal structure and seasonal trends

We run our seasonal radiative-convective 1-D model on 32 distinct columns, each for a different latitude, and for 10 Jupiter years in order to reach radiative-convective equilibrium. We performed three runs corresponding to polar haze scenario #1, 2 and 3. In this section, we mostly present and discuss the results obtained with scenario #2, as we will see in section 3.4 that it appears more consistent with observations. The corresponding latitude-pressure cross-section of the temperature obtained at  $L_s=0^\circ$  with scenario #2 is shown in Figure 10. From low- to mid-latitudes, our model reproduces well the tropopause altitude (100 mbar) and temperature (110 K) reported in previous studies (e.g. Conrath et al., 1998; Fletcher et al., 2016). The stratospheric temperature is nearly isothermal in the range 3–0.1 mbar, where it reaches a maximum of 165 K. Above this level, the temperature decreases with altitude as infrared cooling dominates over solar heating. This is in agreement with Kuroda et al. (2014) who find a 5 K temperature decrease between 0.1 and 0.001 mbar, from 160 to 155 K (ie. overall 5 K colder than our model predictions). At latitudes  $50^\circ$ – $70^\circ$ , stratospheric temperatures are found to be colder than at low latitudes, except in the range 3–30 mbar where the warmer temperatures are due to the absorption of solar light by aerosols. Equilibrium temperatures in this pressure and latitudinal range are strongly

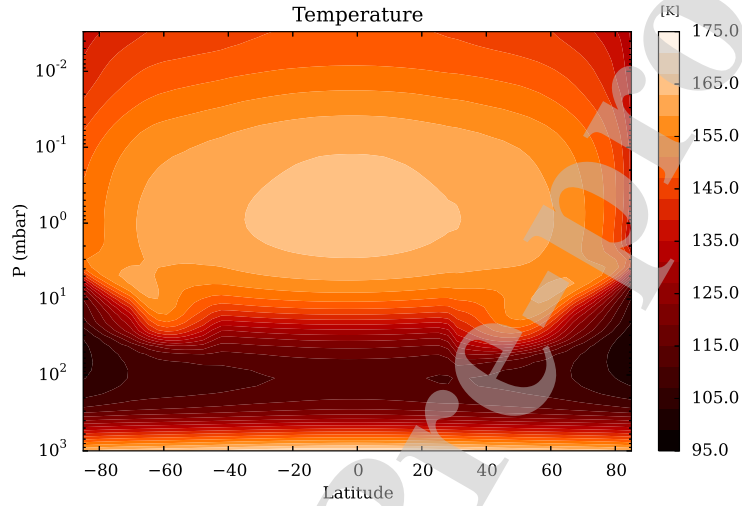


Figure 10: Pressure-latitude cross-section of the temperature (in K) in Jupiter's atmosphere at  $L_s=0^\circ$ . The internal heat flux varies with latitude as defined in eq. 2 and the polar haze scenario #2 was used.

828 influenced by the assumed polar haze properties, as reported in section 2.6.  
 829 However, qualitatively, the thermal structure is similar regardless of the po-  
 830 lar haze scenario. At latitudes poleward of  $70^\circ$ , temperatures are the coldest  
 831 with a 100 K tropopause and a maximum stratospheric temperature of 140 K  
 832 to 150 K at the 1-mbar level. We note that at high latitudes, the tropopause  
 833 is broader and extends from 100 mbar to 20 mbar, which is caused by the  
 834 heating by  $\text{CH}_4$  being less efficient in the lower stratosphere due to the low  
 835 solar elevation.

836 Seasonal variations are expected to be small owing to Jupiter's low obliq-

837 uity. We present in Figure 11 the seasonal evolution of the 10-mbar temper-  
 838 ature at  $60^\circ\text{N}$  and  $60^\circ\text{S}$ , with and without polar haze (scenario #2). We first  
 839 note that the amplitude of seasonal variations is very small in the southern  
 840 hemisphere: it is only 2 K at  $60^\circ\text{S}$ , increasing to 3 K when the radiative  
 841 impact of the polar haze is taken into account. This can be explained by the  
 842 competing effects of obliquity and eccentricity, as Jupiter's perihelion occurs  
 843 at  $\text{Ls}=57^\circ$  close to southern "winter". On the other hand, these two effects  
 844 add up in the northern hemisphere, where the peak-to-peak seasonal ampli-  
 845 tude is  $\sim 6$  K for the polar-haze-free case. When polar aerosols are included  
 846 (scenario #2), there is a global temperature increase of 15 K at  $60^\circ\text{N}$ , 12 K  
 847 at  $60^\circ\text{S}$ . The peak-to-peak amplitude of seasonal variations is also enhanced  
 848 at  $60^\circ\text{N}$  (10 K instead of 6 K) when we include this additional aerosol radia-  
 849 tive forcing. The seasonal amplitude reported here for  $60^\circ\text{N}$  and 10 mbar is  
 850 similar should other pressure levels in the range 30 mbar and 0.01 mbar, and  
 851 latitudes in the range  $45^\circ\text{N} - 75^\circ\text{N}$ , be considered. Finally, at  $60^\circ\text{N}$ , when  
 852 the polar haze are added, we also notice that the temperature maximum is  
 853 shifted to an earlier season ( $\text{Ls}=95^\circ$  instead of  $\text{Ls}=125^\circ$ ), closer to northern  
 854 summer solstice, which hints at shorter radiative timescales as a result of  
 855 adding polar hazes (see next section 3.3 for further details).

### 856 3.3 Radiative timescales

857 In this section, we evaluate and discuss radiative relaxation timescales in  
 858 Jupiter's atmosphere. Radiative timescales can be used to assess whether

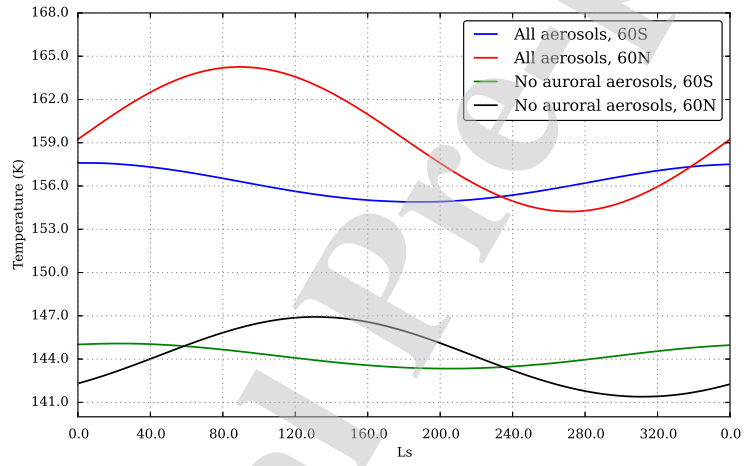


Figure 11: Temperature at the 10 mbar pressure level as a function of solar longitude (Ls, with Ls=0 corresponding to spring equinox in the northern hemisphere) for latitudes 60°N and 60°S, as labeled. Two cases are shown, including or not the stratospheric polar haze (scenario #2).

the atmosphere responds quickly or not to changes in atmospheric temperatures and solar insolation. It is sometimes used in idealized global circulation models where radiative processes are parametrized with a relaxation scheme. Quantitative estimates of the characteristic radiative timescale of the jovian atmosphere have been rather limited in the past, as it requires a detail inventory of the radiative forcings, as is done in this study (see section 2). Recent estimates by Zhang et al. (2013a), Kuroda et al. (2014) and Li et al. (2018a), based on their respective radiative models, take into account gaseous radiative forcings similar to ours, but neglect any kind of clouds and aerosols.

To compute the radiative relaxation timescales of Jupiter’s atmosphere with our seasonal radiative-convective model, we adopt the following standard approach (see e.g. Eq. 6 in Kuroda et al. (2014)): we run a 1-D radiative-convective simulation until radiative equilibrium is reached; then, we add 4 K to the resulting temperature profile at all levels and restart a simulation with this modified profile. Radiative relaxation time,  $\tau_{rad}$ , is obtained by dividing the temperature disturbance (here 4 K) by the change in net (daily-averaged) heating rates due to this disturbance.

Two examples are shown in Figure 12 for latitudes 40°N and the equator. We find that in the upper troposphere, radiative timescales are of the order of 0.2 to 0.4 Jupiter years, meaning that any temperature disturbance due to, for instance, dynamical activity, can persist a long time before being equilibrated by radiative processes. In the stratosphere, this timescale shortens with altitude and is of the order of 3% of a Jupiter year ( $\sim 100$  Earth

days) at the 0.1 mbar level, meaning that a temperature disturbance will be  
 radiatively damped over this timescale (if the source of the disturbance is  
 not active anymore). At the equator, we note that our radiative timescales  
 are in agreement with that derived by Kuroda et al. (2014). Two notable  
 exceptions are the upper stratosphere, where our timescales are about 50%  
 longer than in Kuroda et al. (2014), and the lower troposphere, where our  
 estimated timescale is twice shorter at the 500 mbar level. The former can  
 be explained by the choice of slightly different hydrocarbon profiles at high  
 altitudes and/or differences in spectroscopic calculations, and the latter by  
 the lack of tropospheric aerosols in the model of Kuroda et al. (2014). At  
 40°N and in the range 5–30 mbar, we find that the radiative timescale is two  
 to five times shorter than at the equator. This is due to polar haze radiative  
 forcing (here with scenario #2) and is consistent with our remark on seasonal  
 temperature variations in section 3.2: at the 10-mbar level, the maximum of  
 temperature occurs shortly after summer’s solstice due to a quick response of  
 the atmosphere to changes in solar insolation. This feature was not captured  
 by Kuroda et al. (2014) who neglected the radiative contributions of aerosols  
 in their model. All the conclusions in this paragraph hold when we compare  
 our results to the similar work by Zhang et al. (2013a) and Li et al. (2018a).

### 3.4 Comparison to observations

The monitoring of Jupiter’s spatio-temporal temperature variations from the  
 analysis of thermal infrared spectra started with the Voyager spacecrafts in

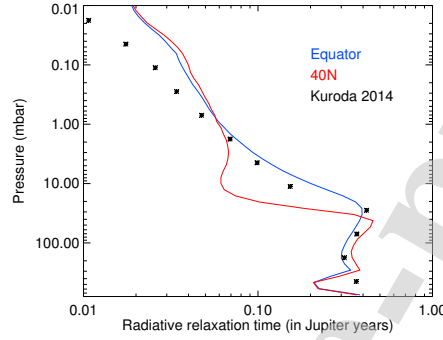


Figure 12: Two example profiles of the radiative timescale in Jupiter's atmosphere at  $L_s=0$  at the equator (in blue) or at  $40^\circ\text{N}$  where polar hazes are abundant (in red, with scenario #2). These are compared to values published in Kuroda et al. (2014) for the equator (stars)

1979 (e.g., Hanel et al., 1979; Simon-Miller et al., 2006) and was later on  
 followed by the Cassini fly-by of Jupiter in December, 2000 (e.g., Flasar  
 et al., 2004; Nixon et al., 2007). Jupiter's thermal structure has also been  
 monitored very regularly from Earth-based telescopes, most notably from  
 NASA's Infrared Telescope Facility (IRTF) (e.g., Orton et al., 1991). Nowa-  
 days, these studies are pursued using the Texas Echelon Cross Echelle Spec-  
 trograph (TEXES) instrument on the IRTF (Lacy et al., 2002; Fletcher et al.,  
 2016; Sinclair et al., 2017; Melin et al., 2018). This high-spectral resolution  
 instrument is able to constrain the 3D temperature field in Jupiter's upper  
 troposphere and stratosphere with a spatial resolution of  $2\text{--}4^\circ$  in latitude  
 (Fletcher et al., 2016), which is actually comparable to the spatial resolution

915 achieved by the Cassini fly-by. Both CIRS and TEXES are sensitive to the  
 916 temperature in the pressure range 700–0.5 mbar (with the caveat of a low  
 917 sensitivity in the 20–60 mbar range) and cover the latitude range  $78^{\circ}\text{S} - 78^{\circ}\text{N}$ .

918 In this section, we focus on the comparison with the results of Fletcher  
 919 et al. (2016) who analysed spectra acquired by TEXES in December, 2014  
 920 (corresponding to  $L_s=175^{\circ}$ ) and also analysed, with the same retrieval pipeline,  
 921 observations from the Composite Infrared Spectrometer (CIRS) on board  
 922 Cassini during the December, 2000 flyby (corresponding to  $L_s=110^{\circ}$ ). In  
 923 our comparisons, we neglect longitudinal variability and only consider tem-  
 924 peratures derived from zonally-averaged spectra provided by Fletcher et al.  
 925 (2016).

926 We first focus on the comparison in the upper troposphere, shown in  
 927 Figure 13. At the 360-mbar level, the temperature derived from CIRS and  
 928 TEXES shows little meridional or temporal variability except at the equator,  
 929 where the TEXES-derived temperature is about 5 K warmer in 2014 than  
 930 the CIRS-derived temperature in 2000. These variations are attributed to  
 931 changes in the dynamics of the equatorial belts (Fletcher et al., 2016). As  
 932 already mentioned in section 2.5.2, the cloudy equatorial zone is colder than  
 933 the less cloudy, warmer equatorial belts at  $15^{\circ}\text{N}$  and  $15^{\circ}\text{S}$ , which is thought  
 934 to be the consequence of vertical motions (upwelling in zones, subsidence in  
 935 belts) rather than due to a radiative effect. Near the tropopause level (at  
 936 110 mbar), observed temperatures exhibit a small (5 K) decrease in temper-  
 937 ature from  $50$  to  $78^{\circ}$  in both hemispheres, and a temporal variability of the



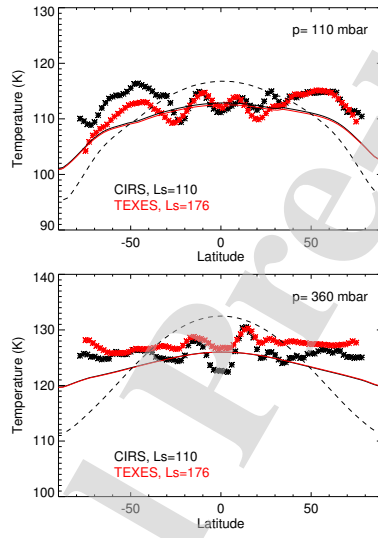


Figure 13: Comparison between tropospheric temperatures derived by Fletcher et al. (2016) from Cassini/CIRS (black stars) and TEXES (red stars) observations at two different seasons, as labeled, and that predicted by our model, in solid lines (in black for  $L_s=110$ , in red for  $L_s=176$ ). The upper and lower panels display temperatures at 110 and 360 mbar, respectively. These results are obtained with a latitudinal-varying internal heat flux; for reference, we also show the simulated temperature obtained when setting a constant internal heat flux (dashed line,  $L_s=110$ ).

938 order of 3 K in the southern hemisphere.

939 Our predicted temperatures, obtained with the parametrization of a latitudinal-  
 940 varying internal heat flux (as is defined in eq. 2) and a single cloud and haze  
 941 scenario, reproduces reasonably well the observed globally-averaged temper-  
 942 ature in the troposphere. Our main disagreement is that at the 360-mbar  
 943 level, our temperatures are  $\sim 4$  K cooler than observations in the  $40\text{--}78^\circ$   
 944 latitude range. The fact that our model still slightly underestimates the  
 945 temperature at high latitudes suggests that we should let the internal heat  
 946 flux increase even more with latitude. However, because this is only a crude  
 947 parametrization that might become obsolete when full GCM simulations are  
 948 run, we did not attempt to optimize the parameters of eq. 2 until we ob-  
 949 tained a perfect match with observations. In addition, other processes could  
 950 be at play in this model-observation mismatch: indeed, the aerosol opacity  
 951 cross-section derived by Zhang et al. (2013b) indicate that the tropospheric  
 952 haze layer could extend at higher altitudes in the  $60\text{--}70^\circ$  latitude range (with  
 953 no observations beyond  $70^\circ$ ), which could enhance the radiative heating in  
 954 the upper troposphere. Given the current lack of observational constraints  
 955 on the haze properties at high latitudes, we did not modify our tropospheric  
 956 haze scenario.

957 Regarding the stratosphere, we present in Figure 14 the meridional tem-  
 958 perature variations at four pressure levels: 0.4, 3, 10 and 25 mbar. Results  
 959 obtained with the three polar haze scenario are shown. Figure 15 compares  
 960 the modeled globally-averaged vertical profiles of temperature, for the three

haze scenario, to that derived from CIRS observations. Several individual vertical profiles of temperature at radiative equilibrium (this time, only for scenario #2 for the sake of clarity) are compared to CIRS and TEXES data at four latitudes in Figure 16. We first note that at all pressure levels, CIRS and TEXES exhibit a strong temporal variability at and near the equator. This region is known to harbor a periodic equatorial quasi-quadriennial oscillation (QJO) in the temperature and associated thermal wind field thought to result from wave-mean zonal flow interactions (Leovy et al., 1991; Orton et al., 1991; Flasar et al., 2004; Simon-Miller et al., 2007), based on analogy with similar oscillations on the Earth and Saturn. Hence, in the following, we will not comment on the model-observation mismatch near the equator, since by design our radiative-convective model cannot capture such a dynamical signature.

At the 10-mbar pressure level, our modeled temperatures exhibit a significant variability depending on the chosen polar haze scenario (see Figure 14). We find that scenario #2 provides an excellent match to the observed temperatures. At this pressure levels, both CIRS and TEXES temperatures feature a local maximum at 50–65°N which is rather well reproduced by our model, should the polar haze scenario #2 be used. If the radiative impact of the polar haze is neglected, the temperature would be 10 to 15 K colder at these pressure levels and latitude range. With the polar haze scenario #1, the comparison with observations is even less favorable, as the haze has a net cooling effect at this pressure level. The hemispherical asymmetry be-

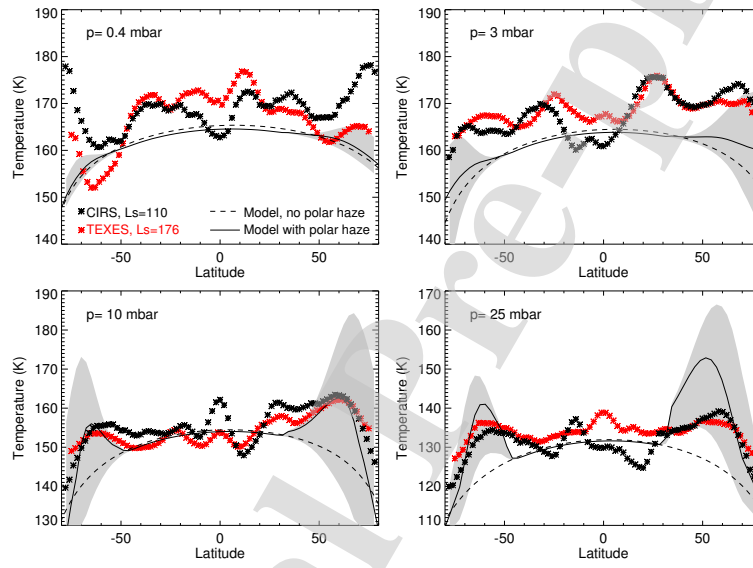


Figure 14: Comparison between stratospheric temperatures derived by Fletcher et al. (2016) from Cassini/CIRS and TEXES observations at two different seasons, as labeled (stars), and that predicted by our model at an intermediate season ( $L_s = 140$ , lines). The dashed line is for a case where the polar haze was neglected while the grey shading represents the effect of including polar haze scenarios #1 to 3 (with the solid black line referring to scenario #2). The four panels display temperatures at four different pressure levels (0.4, 3, 10 and 25 mbar).

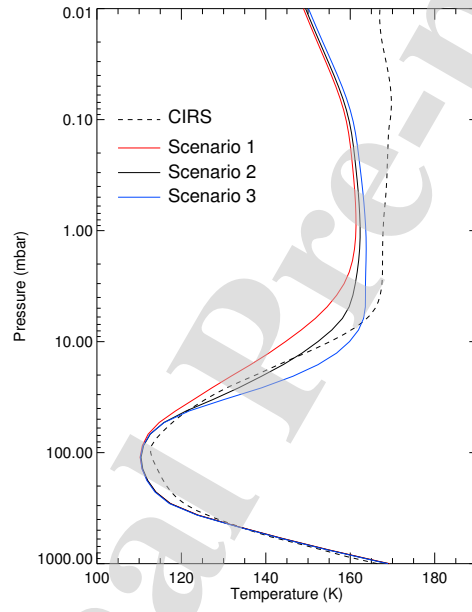


Figure 15: Comparison between temperature vertical profiles averaged between 77°S and 77°N as derived by Fletcher et al. (2016) from Cassini/CIRS observations, in dashed lines, and that predicted by our radiative equilibrium model, in solid lines, for the three polar haze scenario.

984 tween latitudes  $60^{\circ}\text{N}$  and  $60^{\circ}\text{S}$ , of about 8 K, observed by TEXES and CIRS,  
 985 is faithfully reproduced as well with polar haze scenario #2 and #3. We  
 986 can argue that this observed hemispherical asymmetry is caused by a radia-  
 987 tive effect related to the polar haze absorption, as this asymmetry would be  
 988 of only 1–2 K without this radiative contribution. As already discussed in  
 989 section 3.2 and shown in Figure 11, this asymmetry is a seasonal effect: it  
 990 should disappear around  $\text{Ls}=230^{\circ}$  and reverse at  $\text{Ls}=240\text{--}320^{\circ}$  (the tempera-  
 991 ture at  $60^{\circ}\text{S}$  is expected to be 2 K warmer than that at  $60^{\circ}\text{N}$  at that season).  
 992 We note that in their analysis of Voyager observations, shortly after autumn  
 993 equinox ( $\text{Ls}=190^{\circ}$ ), Simon-Miller et al. (2006) found that the temperature  
 994 was  $\sim 6$  K warmer at  $50^{\circ}\text{N}$  compared to  $50^{\circ}\text{S}$  (measurements only extended  
 995 to  $50^{\circ}$  latitude). This is compliant with our results: indeed, due to the strong  
 996 asymmetry in the polar haze as constrained by Zhang et al. (2013), where  
 997 the integrated opacity is found to be about four times greater at  $50^{\circ}\text{N}$  com-  
 998 pared to  $50^{\circ}\text{S}$  (see Figure 8), our model predicts that the temperature at  
 999  $50^{\circ}\text{N}$  remains warmer than that at  $50^{\circ}\text{S}$  throughout the year - provided that  
 1000 the haze hemispheric asymmetry persists.

1001 Our model predicts that a similar north-south asymmetry between  $60^{\circ}\text{N}$   
 1002 and  $60^{\circ}\text{S}$  is still present at the 25-mbar level. This is at odds with CIRS and  
 1003 TEXES observations, which are nearly symmetric about the equator at this  
 1004 pressure level. In addition, the haze scenarios #2 and 3 significantly overesti-  
 1005 mate the high latitude temperatures, especially in the northern hemisphere.  
 1006 On the one hand, this could suggest that we overestimate the aerosol content

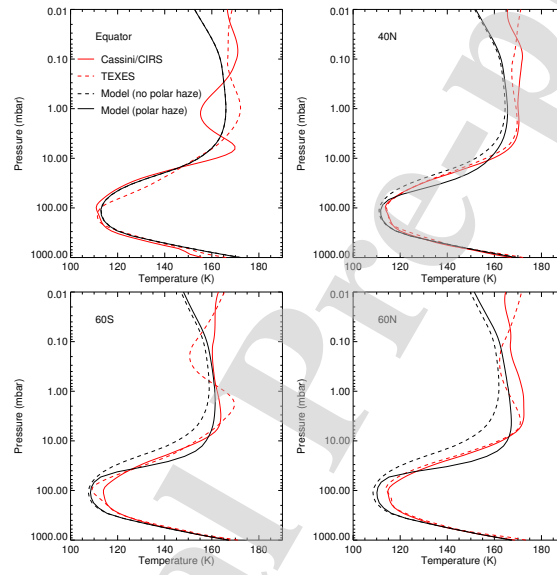


Figure 16: Comparison of vertical profiles of the temperature as derived from our radiative-convective model at four different latitudes (equator, 40°N, 60°S, 60°N), in solid black line, and observed temperature profiles derived from Cassini/CIRS (solid red line) and TEXES (dashed red line) derived by Fletcher et al. (2016). Model results correspond to polar haze scenario #2. For reference, we also show the temperature predicted by our model without the polar haze (dashed black line).

1007 at this pressure level in the northern hemisphere – but not in the southern  
 1008 one, as our predicted temperature agrees better with CIRS and TEXES ob-  
 1009 servations at 50–70°S. The actual vertical profile of aerosol opacity could  
 1010 be different from the one parameterized in eq. 1 and/or be different between  
 1011 60°N and 60°S. In particular, it is important to note that Zhang et al. (2013b)  
 1012 only constrained the shape of the vertical aerosol profile between 25°N and  
 1013 75°S. For latitudes poleward of 25°N, they assumed a similar vertical shape  
 1014 than that at the corresponding southern latitude. Given the large sensitivity  
 1015 of the temperature to this polar haze, it is crucial that future observational  
 1016 studies better characterize its vertical profile in the northern hemisphere as  
 1017 well. On the other hand, we also note that CIRS and TEXES measurements  
 1018 have a rather low sensitivity to the temperature in the 20 to 60 mbar range,  
 1019 so that it is possible that part of the observation-model mismatch at 25 mbar  
 1020 is also due to a larger uncertainty in the observations at this level.

1021 At 10 and 25 mbar, both observations and model predict a sharp decrease  
 1022 in temperature between 65° latitude and the poles. The temperature drop  
 1023 in our model is sharper than the observed one, but this is not surprising:  
 1024 indeed, our model predicts a marked drop due to the sharp decrease of net  
 1025 heating rates at high latitudes. However, it is expected that such a strong  
 1026 temperature gradient would cause dynamical activity (e.g. thermally-direct  
 1027 circulations, baroclinic instability, ...) that would act to counteract this  
 1028 gradient. The study of the associated stratospheric circulation and/or mixing  
 1029 processes is left to a future study.



At lower pressure levels ( $p < 3$  mbar), we note that our predicted temperatures are almost systematically underestimated by  $\sim 5$  K compared to TEXES and CIRS observations, regardless of the chosen polar haze scenario. This is well visible in Fig. 15 on the globally-averaged temperature profiles. This suggests that either a radiative ingredient is missing or not well estimated in our model, or that the temperature is governed by other processes, such as dynamical heating by gravity wave breaking or by eddies. The same conclusion was reached by Zhang et al. (2013a), who discussed these two hypothesis but did not favor one or the other. In a sequel paper, Zhang et al. (2015) find that the atmosphere at pressures  $< 3$  mbar seems to depart from radiative balance (with an excess of radiative cooling at global scale), but the authors emphasize that cooling and heating rate profiles still agree with each other within error bars. It is indeed plausible to reconcile the observed temperature profile (within their  $\sim 2$  K error bars) with our calculated equilibrium temperatures, should the amount of ethane and acetylene be reduced by  $\sim 30\%$  (see sensitivity studies in section 2.4). The typical  $1-\sigma$  error bar on the retrieved abundance of these hydrocarbons is on the order of 20% (Nixon et al., 2010), which makes this scenario plausible, but at the cost of a greater uncertainty on ethane and acetylene mixing ratios than previously thought. Hence, this topic is still an open question.

At 0.4 mbar, we note that TEXES and CIRS observations exhibit important temporal variability at high latitudes (poleward of  $55^\circ$ ). The observed temperature also increases between  $60$  and  $78^\circ$  latitude in both hemispheres,

1053 which is at odds with our simple radiative-convective model. Clearly, other  
 1054 processes must control the temperature at these altitudes. One hypothesis is  
 1055 that the temperature may be influenced by the precipitation of high-energy  
 1056 particles that could warm the atmosphere at high latitudes through Joule  
 1057 heating (Sinclair et al., 2017).

1058 We can also comment on the comparison with Sinclair et al. (2017) who  
 1059 analyzed TEXES and CIRS data specifically focusing on the polar regions.  
 1060 They highlighted a strong local temperature maximum at 1 mbar in Jupiter's  
 1061 auroral oval, which was hypothetically attributed to either Joule heating or  
 1062 absorption by aerosols. In our radiative-convective simulations, the temper-  
 1063 ature maximum associated to aerosol absorption is obtained at 10–20 mbar  
 1064 (and not 1 mbar), where the peak concentration of polar haze is parametrized  
 1065 in our model. However, our simulated conditions are not that of the auroral  
 1066 oval itself: Sinclair et al. (2017) interpretation of strong aerosol heating at  
 1067 1 mbar can hold if there is a local maximum of aerosol absorption at this  
 1068 level in the auroral oval, which remains to be assessed.

1069 This comparison work with state-of-the-art observations shows that our  
 1070 radiative-convective equilibrium model with polar haze scenario #2 repro-  
 1071 duces well, to first order, the observed temperature in the upper troposphere  
 1072 and lower stratosphere ( $p > 5$  mbar), except in the equatorial region (where  
 1073 there is well-known dynamical activity). Other processes might be at play in  
 1074 controlling the temperature in the middle and upper stratosphere, and in the  
 1075 troposphere (belt/zone activity), but the reason behind the systematic  $\sim 5$  K

1076 cold bias at low latitudes in the upper stratosphere is still largely unknown.  
1077 These results are consistent with the work by Zhang et al. (2015), who found  
1078 that the lower and mid stratosphere was near radiative equilibrium should  
1079 a polar haze be included, similar to our haze scenario #2. Using polar haze  
1080 scenario #1 yields too cold temperatures at high latitudes, especially near  
1081 10 mbar, while scenario #3 results in systematically too warm temperatures  
1082 at high latitudes in the 5–30 mbar pressure range. The impact of assuming  
1083 different haze scenario is also well visible at global scale (see Fig. 15). How-  
1084 ever, we caution that the objective of this comparison work is not to “fine  
1085 tune” the haze properties until a match with observations is found, as a local  
1086 radiative imbalance could trigger some atmospheric circulation in the actual  
1087 atmosphere, modifying in turn the temperature. Residual-mean circulations  
1088 induced by such radiative imbalance are estimated in the following section.  
1089 In summary, while the extreme scenario 1 and 3 seem unlikely (given the  
1090 magnitude of the observation-model mismatch and their systematic nature),  
1091 our choosing scenario #2 does not rule out other possible combinations of  
1092 haze properties. The best way forward is to 1) better characterize the polar  
1093 haze in Jupiter’s stratosphere while 2) exploring the stratospheric dynamics  
1094 with the help of a global circulation model for different haze scenarios, and  
1095 confront these future model results back to the observed temperatures.

## 1096 4 Residual-mean stratospheric circulations

1097 In this section, we exploit the computed heating and cooling rates to estimate  
 1098 the residual-mean circulation in Jupiter’s stratosphere. We will in particular  
 1099 explore the impact of assuming different polar haze properties on the residual-  
 1100 mean stratospheric circulation.

### 1101 4.1 Background

1102 Stratospheric circulations are driven by a combination of diabatic and me-  
 1103 chanical (eddy-induced) forcings, resulting in an interplay of transport pro-  
 1104 cesses: advection and mixing. In the Earth stratosphere, it has been shown  
 1105 that the Transformed Eulerian Mean (or residual-mean) circulation is a  
 1106 good approximation to the Lagrangian mean circulation (relevant to tracer  
 1107 transport) in regions where wave breaking and dissipation is relatively weak  
 1108 (Dunkerton, 1978; Butchart, 2014). As we describe in what follows, the  
 1109 residual-mean circulation can be approximately estimated from the knowl-  
 1110 edge of atmospheric net heating rates and temperatures. Hereafter we follow  
 1111 this approach to diagnose the zonally-averaged mass circulation in Jupiter’s  
 1112 stratosphere, for annually-averaged conditions.

1113 The complete equations for the zonally-averaged stratospheric circula-  
 1114 tion are provided by the Transformed Eulerian-mean formulation, with the  
 1115 respectively residual-mean meridional and vertical components of the circu-  
 1116 lation  $(v^*, w^*)$  defined as a combination of a zonal-mean and eddy-induced

1117 terms (Andrews et al., 1987):

$$v^* = \bar{v} - \frac{1}{\rho_0} \frac{\partial}{\partial z} \left( \frac{\rho_0 \overline{v'\theta'}}{\partial\bar{\theta}/\partial z} \right) \quad (4)$$

1118

$$w^* = \bar{w} + \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \left( \frac{\cos \phi \overline{v'\theta'}}{\partial\bar{\theta}/\partial z} \right) \quad (5)$$

1119 where overlines denote zonal averages, primes departures from the zonal mean  
 1120 (eddies),  $\theta$  potential temperature,  $\rho_0$  density,  $a$  planetary radius,  $\phi$  latitude,  
 1121  $z$  altitude. The associated streamfunction  $\Psi$  describing the circulation is  
 1122 defined by

$$(v^*, w^*) = \frac{1}{\rho_0 \cos \phi} \left( -\frac{\partial \Psi}{\partial z}, \frac{1}{a} \frac{\partial \Psi}{\partial \phi} \right) \quad (6)$$

1123 The two components  $(v^*, w^*)$  of the residual-mean circulation follow a mass-  
 1124 conservation equation

$$\frac{1}{a \cos \phi} \frac{\partial (\cos \phi v^*)}{\partial \phi} + \frac{1}{\rho_0} \frac{\partial (\rho_0 w^*)}{\partial z} = 0 \quad (7)$$

1125 and an energy-conservation (thermodynamic) equation

$$\frac{\partial \bar{\theta}}{\partial t} + \frac{v^*}{a} \frac{\partial \bar{\theta}}{\partial \phi} + w^* \frac{\partial \bar{\theta}}{\partial z} = \mathcal{Q} + \mathcal{E} \quad (8)$$

1126 in which  $\mathcal{Q}$  is the net radiative heating rate and  $\mathcal{E}$  is the heating rate related  
 1127 to eddies forcing the mean flow

$$\mathcal{E} = -\frac{1}{\rho_0} \frac{\partial}{\partial z} \left[ \rho_0 \left( \frac{\overline{v'\theta'}}{a} \frac{\partial \bar{\theta}}{\partial \phi} + \overline{w'\theta'} \right) \right] \quad (9)$$

For quasi-geostrophic large-scale flows in non-acceleration conditions, the eddy-related term  $\mathcal{E}$  can be neglected. Under this approximation, the residual-mean circulation is similar to the so-called diabatic circulation also used on Earth to diagnose the Brewer-Dobson circulation (Butchart, 2014). Additionally, considering atmospheric state and circulation averaged over a year, the temporal term in eq. 8 can also be neglected, which entails

$$\frac{v^*}{a} \frac{\partial \bar{\theta}}{\partial \phi} + w^* \frac{\partial \bar{\theta}}{\partial z} \simeq \mathcal{Q} \quad (10)$$

It is important to note here that neither the seasonal variations of temperature nor the impact of eddies are negligible in Jupiter's stratosphere; yet the approximations are reasonable in a context where we seek the average meridional and vertical transport experienced by the long-lifetime chemical species in Jupiter's stratosphere.

## 4.2 Method

Equations 7 and 10 are solved to obtain the residual-mean circulation  $(v^*, w^*)$  under the approximations  $\mathcal{E} \simeq 0$  (we neglect the eddy heat flux convergence term) and  $\partial \bar{\theta} / \partial t \simeq 0$  (we neglect seasonal variations). In equation 10, the temperature profiles  $\bar{\theta}(z)$  are an averaged of Cassini/CIRS and IRTF/TEXES observations analyzed by Fletcher et al. (2017). The choice of averaging these two data sets is motivated by the will to smooth out a bit the large amplitude of the QQQ signal and to get a better representation of a seasonally aver-

aged temperature at  $L_s \sim 140^\circ$ . We compute the net radiative heating rates  $\mathcal{Q}$  by running our seasonal radiative model for just one time step, starting at  $L_s = 140^\circ$  and with a temperature field corresponding to the observed temperatures interpolated on our model grid. We repeat this work for the three polar haze scenario described in section 2.6 to test the sensitivity of the diabatic circulation to these different radiative forcings.

To solve equations 7 and 10, we use the iterative method described in Solomon et al. (1986) :

1. At the initial iteration  $i = 0$ , the meridional component  $v_{i=0}^*$  is set to zero in equation 10, and we simply solve for the vertical component  $w_{i=1}^*$  given the vertical gradient of potential temperature (as if simply computing adiabatic warming/cooling by subsiding/ascending motions equilibrating the radiative heating rate).
2. The vertical component  $w_{i=1}^*$  is used to obtain the streamfunction  $\Psi_{i=1}$  by integrating equation 6 (using a Simpson integration method)

$$\Psi_{i=1} = \int_{-\frac{\pi}{2}}^{\phi} (w_{i=1}^* + \epsilon) \cos \phi \, a \, d\phi \quad (11)$$

where

$$\epsilon = \left( \int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} \cos \phi \, d\phi \right)^{-1} \int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} w_{i=1}^* \cos \phi \, d\phi \quad (12)$$

is a corrective term (usually a couple percent at best) designed to ensure that the streamfunction  $\Psi_i$  vanishes at the north pole.

- 1165 3. The meridional component  $v_{i=1}^*$  is obtained from the streamfunction  $\Psi_{i=1}$   
 1166 by using equation 6 (vertical derivative of  $\Psi_{i=1}$ ). Using the stream-  
 1167 function  $\Psi_{i=1}$  to compute  $v_{i=1}^*$  from  $w_{i=1}^*$  is equivalent to solving the  
 1168 mass-conserving equation 7.
- 1169 4. The meridional component  $v_{i=1}^*$  is injected in equation 10 to obtain  
 1170 the vertical component  $w_{i=2}^*$  at the next iteration, then the process is  
 1171 looped back to step 2 for  $i = 2$ .

1172 This iterative procedure converges quickly: iterations  $i > 3$  yield a change  
 1173 from  $(v_i^*, w_i^*)$  to  $(v_{i+1}^*, w_{i+1}^*)$  of about 1 %. We stopped the computations  
 1174 at the tenth iteration in which the increment from the previous step is only  
 1175 0.01 %. Our algorithm was checked upon a well-constrained analytical ex-  
 1176 ample.

### 1177 4.3 Results and comparison to previous studies

1178 Hereafter, we mostly describe the results obtained with the most favorable  
 1179 (according to section 3) polar haze scenario #2. The resulting streamlines of  
 1180 the residual circulation, for the stratosphere only, are displayed in Figure 17  
 1181 and the corresponding vertical and meridional wind speeds are shown in Fig-  
 1182 ure 18. For reference, the pressure-latitude cross section of the net radiative  
 1183 heating rates  $\mathcal{Q}$  used to derive this circulation is shown in Figure 19.

1184 Overall, many small circulation cells are present, due to the fact that there  
 1185 are many local extrema in the spatial distribution of net heating rates  $\mathcal{Q}$ .



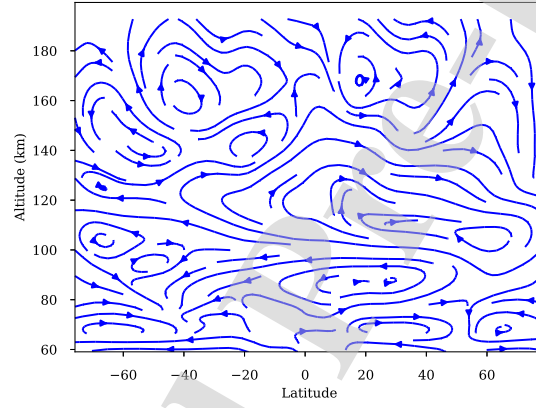


Figure 17: Streamlines computed from eq. 11, using the polar haze scenario #2 and the averaged temperature derived from Cassini/CIRS and TEXES. The altitude is computed with the convention  $z=0$  km at the 1-bar level. For reference, the bottom of the figure, at 50 km, corresponds to the lower stratosphere ( $\sim 50$  mbar), while the 1 mbar level lies at  $\sim 135$  km. For the sake of clarity, the vertical component has been multiplied by 900 in this figure since the horizontal scale from one pole to the other, in km, is  $\sim 900$  times the vertical extent considered here (140 km).

1186 Nevertheless, two prominent large-scale circulation cells can be noted. In the  
 1187 lower stratosphere (10 to 30 mbar, or  $\sim 80\text{--}110$  km altitude), the residual-  
 1188 mean circulation is characterized by upwelling at  $50\text{--}60^\circ\text{N}$  and cross-equatorial  
 1189 flow from northern high latitudes to southern mid latitudes. The vertical  
 1190 wind speed at  $50^\circ\text{N}$  reaches  $0.12\text{ mm.s}^{-1}$  at the 20-mbar level and the merid-  
 1191 ional wind speed is of the order of  $0.10\text{--}0.15\text{ m.s}^{-1}$ . This cell is forced by  
 1192 the net positive heating rate centered at  $50^\circ\text{N}$  at a pressure level of 20 mbar.  
 1193 Indeed, as discussed in section 3.4, the expected equilibrium temperature at  
 1194 this location is much warmer than that observed by CIRS and TEXES. Sub-  
 1195 sequently, when using the observed temperatures to diagnose the circulation,  
 1196 an upwelling is needed to balance the "too cold" observed temperature. In  
 1197 other words, the net radiative heating is compensated by a diabatic cooling.  
 1198 Should the polar haze scenario #1 be used (characterized by net negative  
 1199 heating rate near  $50\text{--}70^\circ\text{N}$ , 3–20 mbar), the circulation would reverse, with a  
 1200 strong downwelling occurring at high latitudes near 5–10 mbar and equator-  
 1201 to-pole meridional wind centered at 5 mbar, in both hemispheres. The sen-  
 1202 sitivity of the diabatic circulation to the polar haze scenario is illustrated in  
 1203 Figure 20, which shows the vertical wind  $w^*$  at the 10-mbar pressure level  
 1204 for each of the three haze scenario. This illustrates well that accurate knowl-  
 1205 edge of radiative forcings is crucial to employ this method to understand the  
 1206 jovian stratospheric circulation. By extension, this knowledge will also be  
 1207 crucial for future GCM simulations.

1208 In the middle stratosphere ( $p < 3$  mbar or  $z > 120$  km), the circulation

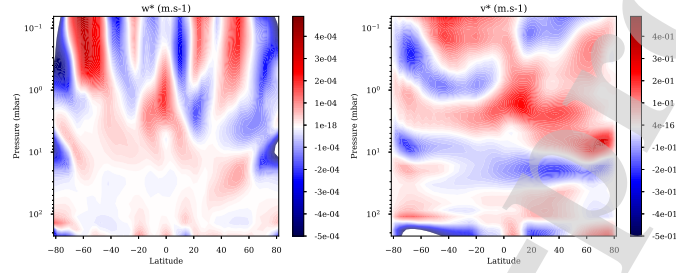


Figure 18: Pressure-latitude cross-section of the vertical (left) and meridional (right) components of the residual-mean circulation, in  $\text{m.s}^{-1}$ , estimated using the polar haze scenario #2 and temperatures derived from Cassini/CIRS and TEXES.

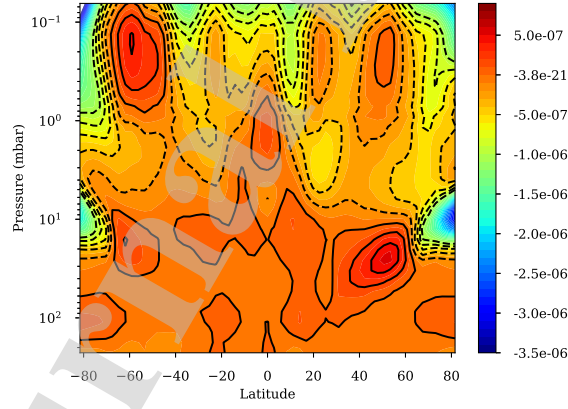


Figure 19: Pressure-latitude cross-section of the net heating rates in Jupiter's stratosphere, in  $\text{K.s}^{-1}$ , estimated using the polar haze scenario #2 and temperatures derived from Cassini/CIRS and TEXES.

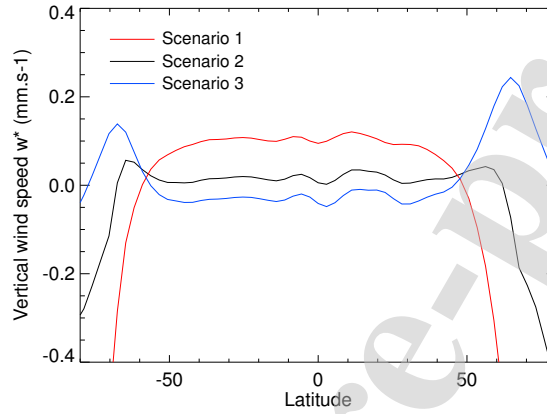


Figure 20: Vertical component of the residual-mean circulation at the 20-mbar pressure level, in  $\text{mm.s}^{-1}$ , estimated using the polar haze scenario #1, 2 or 3, as labeled.

1209 does not depend on the chosen polar haze scenario. The diabatic circulation  
 1210 there is dominated by upwelling in the  $20^{\circ}\text{S}$ – $0^{\circ}$  region with cross-equatorial  
 1211 meridional flow and subsidence poleward of  $50^{\circ}\text{N}$ . Vertical wind speed at the  
 1212 equator reaches  $0.25 \text{ mm.s}^{-1}$  at the 2-mbar pressure level while the meridional  
 1213 wind speed in the northern hemisphere is of the order of  $0.2$ – $0.3 \text{ m.s}^{-1}$ . The  
 1214 significant upwelling motion near the equator and 1 mbar is clearly associ-  
 1215 ated with the thermal structure of the Quasi-Quadriennial Oscillation (QQO)  
 1216 and its particular phase at the time of Cassini/CIRS observations in 2000  
 1217 (see figure 16). Averaging CIRS temperature profiles with the ones retrieved  
 1218 from TEXES smooths out partially this feature, but not entirely. The ob-  
 1219 served temperature field exhibits a local minimum near 1 mbar (figure 14),

1220 which translates into upwelling (hence, diabatic cooling) when diagnosing  
 1221 the circulation. Given the periodic nature of the QQO signal, this feature is  
 1222 unlikely to be a part of the annually-averaged meridional circulation.

1223 The fact that the dominant meridional motion is from equator to high  
 1224 northern latitudes, with stronger downwelling at high northern than at high  
 1225 southern latitudes, stems from a hemispheric asymmetry in the net heating  
 1226 rates. This asymmetry itself results from an asymmetry in the observed  
 1227 temperature at high latitudes (see figure 14). However, it is unclear whether  
 1228 this asymmetry reflects an overall seasonal effect or more transient conditions.  
 1229 In addition, observed departures from the expected equilibrium temperature  
 1230 in polar regions might be caused by processes (Joule heating, etc) other  
 1231 than the dynamical motions that are diagnosed with this method. Another  
 1232 potential issue is related to the systematic underestimation, by our radiative  
 1233 equilibrium model, of the observed temperature by  $\sim 5\text{K}$  at  $p < 3$  mbar, as  
 1234 discussed in section 3.4. Energy transfer by eddies could be important at  
 1235 these pressure levels and is one possible explanation of this mismatch. If that  
 1236 is the case, then neglecting  $\mathcal{E}$  in equation 8 is not valid and using the observed  
 1237 (warmer) temperatures for computing the cooling rates while neglecting  $\mathcal{E}$   
 1238 when estimating the streamfunction could lead to erroneous results. Hence,  
 1239 this is another example of the limitations of this diabatic circulation method,  
 1240 which requires a high degree of confidence on both the observed temperature  
 1241 field and its drivers.

1242 Previous estimates of the residual-mean circulation in Jupiter's atmo-

sphere were obtained by West et al. (1992) and Moreno and Sedano (1997) using a similar formalism than ours, albeit based on the temperature field derived by Voyager/IRIS to compute the cooling rates. These temperature profiles had a low vertical resolution in the stratosphere and were provided at only two pressure levels: 77 mbar and 1 mbar. For intermediate pressure levels, West et al. (1992) and Moreno and Sedano (1997) had to rely on interpolation. Hence, one strong limitation to their study is that the observational constraint they used (the Voyager/IRIS temperature fields) did not capture the polar haze region, mainly located near 10–30 mbar. West et al. (1992) noticed that using this temperature field resulted in an imbalance between cooling and heating rates at global scale, with residual net radiative heating at the 10-mbar pressure level. While the authors did include the effect of absorption of UV and visible light by a stratospheric polar haze for the computation of solar heating rates, they most probably underestimated the cooling rates due to a lack of detailed knowledge of the thermal structure.

To mitigate this issue, West et al. (1992) scaled either the solar heating rate, or the infrared cooling rate, by a factor that depended on altitude only. Moreno and Sedano (1997) noted a similar issue and chose to scale the temperature profile with a factor that depended on height but not with latitude, until radiative balance was achieved at global scale. We can question these choices, as the resulting cooling rate would most certainly be missing a latitudinal-varying term (linked to the meridional distribution of the stratospheric haze and its impact on the thermal structure). In addition,

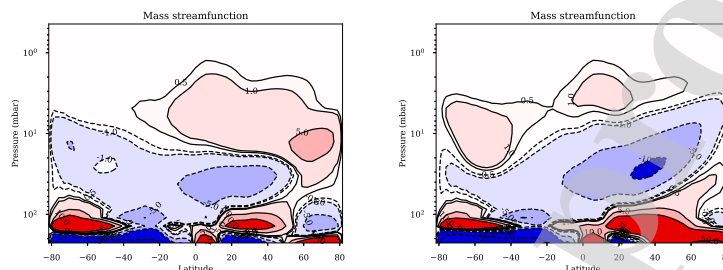


Figure 21: Mass streamfunction, in units of  $\text{kg} \cdot \text{m}^{-1} \cdot \text{s}^{-1}$ , computed from eq. 11 using either the polar haze scenario #2 (left panel) or scenario #3 (right panel). Results are shown up to 0.3 mbar to facilitate comparison with previous work by West et al. (1992) (note that the latter used different units, in  $\text{g} \cdot \text{m}^{-1} \cdot \text{s}^{-1}$ , so that a factor of 10 exists between the numerical values in this figure and fig. 7c of West et al. (1992)).

neither study included the contribution from the polar haze when computing the cooling rates; they only took into account gaseous contributions, which is now known to be largely inaccurate (Zhang et al., 2015). Hence, for these reasons, the circulations derived by West et al. (1992) and Moreno and Sedano (1997) were probably flawed by these shortcomings.

Nevertheless, there are similarities with the circulation derived in this paper. The diabatic circulation estimated by West et al. (1992) is characterized by two cells with upwelling branches centered at 60°N and 70°S and at a pressure level of 3–10 mbar, driven by significant net radiative heating rates at these locations. Maximum vertical wind speeds of the order of  $10^{-4}$  m.s<sup>-1</sup> (shown in Fig. 4 of Friedson et al. (1999)) are reached near 5 mbar. The dominant meridional flow in West et al. (1992) is from high latitudes to low

latitudes in the 3–30 mbar range, with subsidence occurring over a broad tropical region. In our derived circulation with the reference scenario #2, an upwelling branch with similar vertical wind speed is present in the northern hemisphere, although located slightly deeper, near the 20-mbar level. Its counterpart in the southern hemisphere is much weaker. This difference with West et al. (1992) likely results from differences in the vertical distribution of haze and/or in the vertical temperature profile, as argued above. However, considering the polar haze scenario #3 (with more absorbant haze), upwelling branches are present in both hemispheres, similarly to that derived by West et al. (1992). To illustrate this effect and facilitate the comparison with previously published work, we show in Figure 21 the mass streamfunctions derived when assuming either scenario #2 or #3 that can be compared with Fig. 7(c) of West et al. (1992). The mass streamfunction obtained with scenario #3 is actually very similar to that derived by West et al. (1992), both qualitatively and quantitatively. This is not surprising: as discussed above, we suspect that West et al. (1992) analysis underestimated locally the actual cooling rates at the location where the polar haze warms the atmosphere. This resulted in large net heating rates at 60–70°S. Our scenario #3 exhibits large net heating rates as well, although for other reasons (due to greater absorption by the polar haze). This explains the similarity in the circulation derived in both these studies. This represents another illustration of the sensitivity of the results to the assumed haze properties and to the observed temperature field used to compute cooling rates.



1301 Finally, we can compare our derived residual vertical wind speeds with  
 1302 that of the eddy diffusion coefficient  $K_{zz}$  taken from models A, B and C  
 1303 of Moses et al. (2005). In the range 1 to 30 mbar, estimates of  $K_{zz}$  are of  
 1304 the order of  $2 \times 10^3$  to  $10^4$   $\text{cm}^2.\text{s}^{-1}$ . An order of magnitude of the vertical  
 1305 velocity induced by eddy motions can be obtained by dividing  $K_{zz}$  by the  
 1306 atmospheric scale height (25 km). This results in vertical wind speeds linked  
 1307 with eddy diffusivity of 1 to  $4 \times 10^{-5}$   $\text{m.s}^{-1}$ , *ie.* just half the typical values  
 1308 of  $w^*$  obtained in our study.

#### 1309 4.4 Implications for the transport of trace species

1310 As discussed above, the circulation derived by our study – even though us-  
 1311 ing more detailed temperature fields and state-of-the-art opacities – suffers  
 1312 from significant uncertainties as well. Given this uncertainty, we did not at-  
 1313 tempt to evaluate the impact of this circulation on the distribution of trace  
 1314 species. However, we can comment on several existing studies, based on  
 1315 the lessons learned from this exercise. For instance, Hue et al. (2018) study  
 1316 the distribution of ethane and acetylene, the main by-products of methane  
 1317 photochemistry, with the goal of explaining the observed increase in ethane  
 1318 towards high latitudes, while acetylene is decreasing. The authors combine  
 1319 a photochemical model with a simple parametrization of transport (both ad-  
 1320 vection and diffusion). In doing so, they test several Hadley-like circulation  
 1321 cells with upwelling at the equator and subsidence at both poles. The authors  
 1322 fail to explain the opposed distributions of acetylene and ethane. Our study

1323 suggests that assuming such circulation cells is not appropriate in the lower  
 1324 stratosphere, where these puzzling hydrocarbon distributions are observed.  
 1325 More complex circulation patterns, such as those shown in Figure 21, are  
 1326 probably at play and need to be further understood.

1327 Another puzzle is related to  $\text{CO}_2$  and HCN, products of the SL-9 im-  
 1328 pact that occurred at  $44^\circ\text{S}$ , which display opposite trends several years after  
 1329 the impact ( $\text{CO}_2$  being maximum at the south pole while HCN is found  
 1330 well-mixed from mid-southern to mid-northern latitudes). One hypothesis  
 1331 proposed by Lellouch et al. (2006) is that HCN and  $\text{CO}_2$  were deposited at  
 1332 different altitudes and were transported by different wind regimes. If we as-  
 1333 sume that the diabatic circulation derived from the polar haze scenario #2 is  
 1334 realistic, then this scenario implies that HCN was deposited near 0.5–5 mbar,  
 1335 to be transported equatorward, and  $\text{CO}_2$  was deposited at pressures either  
 1336 lower than 0.5 mbar or around 10 mbar to be transported poleward (see  
 1337 Figure 18). Such scenario need to be tested in the future with chemistry-  
 1338 transport models.

## 1339 5 Conclusions

1340 We have developed a radiative-convective equilibrium model for Jupiter's  
 1341 troposphere and stratosphere that includes parametrizations of several cloud  
 1342 and haze layers. As for its Saturn counterpart (Guerlet et al., 2014), this  
 1343 model is computationally efficient and aims at being coupled with a dynam-

1344 ical core of a General Circulation Model (as was recently done in the Saturn  
1345 case, see Spiga et al., 2020). We take into account the radiative contribution  
1346 of :

- 1347 •  $\text{CH}_4$ ,  $\text{C}_2\text{H}_6$ ,  $\text{C}_2\text{H}_2$  and  $\text{NH}_3$  for radiatively active species along with  
1348 collision-induced absorption from  $\text{H}_2\text{--H}_2$ ,  $\text{H}_2\text{--He}$ ;
- 1349 • A rather compact ammonia cloud located at 840 mbar comprising 10-  
1350  $\mu\text{m}$  particles, with a visible integrated opacity of 10.
- 1351 • A tropospheric haze layer extending between 660 and 180 mbar com-  
1352 posed of 0.5  $\mu\text{m}$  particles with near "grey" optical constants and an  
1353 integrated opacity of 4.
- 1354 • A stratospheric haze layer made of fractal aggregates (typically, 1000  
1355 monomers of 10-nm each, fractal dimension of 2) supposedly linked with  
1356 precipitation of high-energy particles at high latitudes. Their opacity  
1357 is maximum near the 20-mbar level and at latitudes poleward of 30°N  
1358 and 45°S.

1359 The gaseous abundance profiles as well as tropospheric cloud and haze  
1360 layer properties are fixed in latitude and time in the current version of our  
1361 model, but we have studied the sensitivity to varying those parameters. For  
1362 instance, varying the abundance of hydrocarbons with latitude in similar  
1363 proportions than the observed ones (i.e. a poleward enhancement in ethane  
1364 by a factor of two, while acetylene is reduced by 50%) lead to temperature

changes of at most 4 K in the 1 to 0.1-mbar level. Increasing the cloud opacity by a factor of two yields a temperature increase of 3 K in the upper troposphere. These changes are rather small and our main conclusions are not hampered by these simplistic assumptions. The inclusion of photochemistry (that would compute realistic hydrocarbon variations) or cloud microphysics (that would simulate spatio-temporal evolution of cloud formation on Jupiter's atmosphere) is devoted to a future study.

We confirm that the stratospheric polar aerosols have an important role in the radiative budget of Jupiter's stratosphere, yet with a significant uncertainty regarding its magnitude. Their net impact at latitudes 45–60° is maximum in the range 5–30 mbar and depends on their assumed properties (refractive index, size and number of monomers). A large contribution of aerosols to the heating rates was already demonstrated by Zhang et al. (2015) (and before that by West et al., 1992), but it is the first time that the impact of aerosols on stratospheric temperatures is studied. We tested the response of the atmosphere (in terms of radiative equilibrium temperatures) to three different polar haze scenario and find that the reference model of Zhang et al. (2013b) provides a satisfactory comparison to observations at the time of the Cassini flyby in 2000 ( $L_s=110^\circ$ ) or to a TEXES observing run in 2014 ( $L_s=176^\circ$ ). The other two models tend to either systematically overestimate or underestimate the observed temperature in the 10–30 mbar range.

We find that the hemispheric asymmetry in stratospheric aerosols opacity

(that is much larger in the northern than in the southern hemisphere) combined with the small obliquity and eccentricity of Jupiter cause the predicted 10-mbar temperature to be systematically warmer at 50°N than at 50°S, throughout the year. The asymmetry of 8 K in temperature at the 10-mbar level reported by Fletcher et al. (2016) between 60°N and 60°S is also well reproduced by our model with polar haze scenario #2. We also find that the polar haze significantly shortens the radiative timescales, estimated in section 3.3 to 100 days, or 3% of a Jupiter year, at the 10-mbar level. Nevertheless, significant uncertainties remain regarding the optical properties, sizes, meridional distribution and temporal variations of this stratospheric polar haze. This prevents an advanced interpretation of the model-observations mismatch.

At lower pressures ( $p < 3$  mbar), we find that the modeled temperature is systematically lower than the observed one, by typically 5 K. This is consistent with the previous study of radiative budgets in giant planet atmospheres by Li et al. (2018a), who find that the cooling rate exceeds the heating rate in a large part of Jupiter's stratosphere, and with previous studies by Zhang et al. (2013a) and Kuroda et al. (2014). However, as already noted by Zhang et al. (2015), radiative equilibrium could be reached considering the uncertainties on heating and cooling rates associated with the uncertainty on the abundance of hydrocarbons. In our case, warmer equilibrium temperatures can be reached if we assume that ethane and acetylene are currently overestimated by roughly 30%. Other possible explanations are: a missing ra-

1411 diative ingredient ; a mechanical forcing (such as gravity wave breaking or  
 1412 other eddy terms warming the atmosphere) ; a coupling with thermospheric  
 1413 or ionospheric circulations, through Joule heating, adiabatic compression or  
 1414 horizontal advection (e.g. Majeed et al., 2005, although the upper strato-  
 1415 sphere is marginally covered in their study). These scenario need to be further  
 1416 evaluated.

1417 In theory, and under the assumption that the eddy heat flux convergence  
 1418 term is negligible, the knowledge of net radiative heating rates can be then  
 1419 exploited to estimate the stratospheric residual-mean circulation, in a similar  
 1420 fashion as West et al. (1992). In the Earth's stratosphere, the residual-mean  
 1421 circulation represents well, on a seasonal scale, the transport of tracers in  
 1422 regions where wave breaking and dissipation are weak (Butchart, 2014, and  
 1423 references therein). This topic is of high interest on Jupiter, as the observed  
 1424 meridional distribution of photochemical products (ethane and acetylene) or  
 1425 by-products of comet Shoemaker Levy 9 impact (HCN, CO<sub>2</sub>, dust...) is puz-  
 1426 zling and cannot be explained by simple chemistry-transport models. We  
 1427 revisited the study by West et al. (1992) and Moreno and Sedano (1997) by  
 1428 estimating the diabatic circulation based on state-of-the-art knowledge on  
 1429 opacity sources and observed temperature fields from the Cassini fly-by of  
 1430 Jupiter in December 2000 and ground-based TEXES observations in 2014.  
 1431 Our main conclusion is that our current limited knowledge on the differ-  
 1432 ent radiative forcing terms (in particular regarding the stratospheric haze  
 1433 properties) and mechanical forcings (related to the magnitude of eddy heat

flux) results in a low-to-moderate confidence in the estimated circulation. On Earth, both the temperature field and the net radiative heating rates are known with a much higher degree of confidence, allowing to derive the main circulation patterns from this method. The lessons learned from trying to adapt this method to Jupiter are that more investigations are needed regarding the characterization of Jupiter's polar haze radiative properties and other drivers of the meridional circulation. Observations of the polar regions are challenging from Earth, but feasible. The Juno spacecraft offers unique views of the poles and could help characterizing the haze in the UV and near infrared. The future JUICE mission will also include a specific science phase at relatively high obliquity to get good views of Jupiter's polar regions, from the UV to  $5\mu\text{m}$ .

To conclude, we have documented here the building and validation of a radiative-convective model for Jupiter, discussed the resulting equilibrium temperature, how it compares with observations and attempted to derive the residual-mean circulation associated with the computed net heating rates. In order to go further into understanding Jupiter's atmospheric circulations, both in the troposphere and stratosphere, current efforts are focused on running 3D GCM simulations for Jupiter using a hydrodynamical solver coupled with the radiative seasonal model described herein. This will give insights into – among other topics – understanding what is the role of eddies in controlling the stratospheric circulations, mixing and thermal structure ; what governs the distribution of trace species and what are the mechanisms driv-

1457 ing the QO. These topics are also valid for Saturn's atmosphere, which  
1458 shares similar open questions regarding its atmospheric circulation but with  
1459 different seasonal forcings, opening the way to comparative studies between  
1460 these two gas giants.

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## Data availability

The 3-D (latitude, pressure, time) temperature field for the reference radiative-convective simulation will be made available, in the form of a NetCDF file with a doi, on the data service hosted by Institut Pierre Simon Laplace when the paper will be accepted.

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Highlights :

A seasonal radiative-convective model of Jupiter's atmosphere is presented.  
Stratospheric polar haze greatly impact the equilibrium temperatures.  
We evaluate the residual-mean stratospheric circulations and discuss caveats.