Radiative-equilibrium model of Jupiter’s atmosphere and application to estimating stratospheric circulations

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July 11, 2019

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Abstract

Jupiter’s upper troposphere and stratosphere is host to a rich dynamical and chemical activity, modulating 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 radiative-convective model of Jupiter’s atmosphere taking into account radiative forcings from the main hydrocarbons (methane, ethane, acetylene), ammonia, collision-induced absorption, several tropospheric and stratospheric cloud and haze layers and an internal heat flux. Sensitivity studies to several parameters (hydrocarbon abundances, cloud particle sizes and composition, ...) are presented, and we discuss the expected seasonal, vertical and meridional temperature variations in Jupiter’s atmosphere. We then compare the resulting thermal structure with that derived from thermal infrared observations by Fletcher et al., 2016. One of our main findings is that the modeled temperature is in excellent agreement with observations at and near the 10-mbar pressure level, where auroral aerosols (set to follow closely the spatial distribution derived by Zhang et al., 2013) warm the atmosphere by up to 20K at latitudes 45–60°. Auroral aerosols also act to shorten significantly the radiative timescales, estimated to 100 (Earth) days at the 10-mbar level. At pressures lower than 3 mbar, our modeled temperatures are systematically underestimated by ∼5K compared to observations, suggesting that other processes, such as dynamical heating by gravity wave breaking or by eddies, or a coupling with thermospheric circulation, play an important role. Regarding the troposphere, we can only match the observed lack of meridional gradient of temperature by varying the internal heat flux with latitude. In a second part of this paper, we exploit knowledge of heating and cooling rates to diagnose the residual-mean circulation in Jupiter’s stratosphere, under the assumption that the eddy heat flux convergence term is negligible, in a similar fashion as West et al., 1992. In the lower stratosphere (1–30 mbar), the residual-mean circulation is characterized by upwelling at high latitudes with vertical velocities of 0.3 mm.s⁻¹ (due to net radiative heating by the polar stratospheric haze) and subsidence at 10°S. In the middle and upper stratosphere, the circulation turns into Hadley-like circulation cells with rising motion at the equator and subsidence at both poles. Future work will focus on running 3D simulations with a global climate model and evaluate in particular eddy-forcing terms and their influence on stratospheric circulation and temperature.

Highlights

• A radiative-convective model of Jupiter’s atmosphere is presented
• Stratospheric haze greatly impact temperatures, in agreement with observations
• We evaluate the residual-mean stratospheric circulation
1 Introduction

In the past decades, infrared spectroscopy and imaging from instruments onboard the Pioneer, Galileo, Voyager, Cassini and Juno spacecrafts, together with ground-based telescopes, have revealed in great detail the circulations, thermal structure and composition variations in Jupiter’s upper troposphere and stratosphere. These observations give insights into major radiative, chemical and dynamical processes at play in the jovian atmosphere; yet several puzzling features remain hitherto unexplained, particularly in the stratosphere.

In the middle stratosphere (1-10 hPa), the observed meridional distributions of ethane and acetylene – by-products of the methane photochemistry – are found to be at odds with the predictions of one-dimensional (1-D) neutral photochemical models. While acetylene is maximum at low latitudes, following the yearly-averaged insolation as is expected from photochemistry, long-lived ethane increases towards the poles (e.g., Nixon et al., 2007). To explain this trend, redistribution by meridional circulation was suggested to occur on timescales greater than acetylene’s lifetime (∼10 years) and shorter than ethane’s (∼500 years). However, the existing models coupling neutral photochemistry and circulation (with a simple parametrization of meridional and vertical diffusion and advection) have failed so far to explain the opposite distributions of acetylene and ethane (Hue et al., 2018). A possible explanation is given by Sinclair et al. (2017) who suggest that ion chemistry locally enhances the production of acetylene in the auroral oval (where high energy electrons and ion precipitates). Acetylene is then converted into ethane out of the auroral oval, providing a possible explanation for the net increase in ethane at high latitudes.

Another puzzling observation is related to comet Shoemaker-Levy 9 (SL 9) debris in Jupiter’s atmosphere and their migration several years after the 1994 impact. Observations of the spreading of dust in the 1–200 mbar region (Friedson et al., 1999), as well as HCN, CS, CO and CO₂ in the 0.1-0.5 mbar region (Moreno et al., 2003; Griffith et al., 2004; Lellouch et al., 2006; Cavalié et al., 2017), can be used to trace stratospheric circulations. One of the most striking results is that, 6.5 years following the impact, HCN was found to be efficiently mixed from the impact site (44°S) to northern mid-latitudes while CO₂ was found greatly enhanced near the south pole (Lellouch et al., 2006). On the one hand, the authors could match the observed HCN distribution with a model combining equatorward advection with a meridional eddy mixing coefficient reaching a maximum at the equator and decreasing toward high latitudes. On the other hand, a constant eddy mixing coefficient coupled to poleward advection was needed to account for CO₂ enhancement near the south pole. To reconcile these different trends, Lellouch et al. (2006) suggested that CO₂ and HCN lie at different altitude levels, acknowledging that this represents another puzzle since both species are thought to be deposited by SL-9 or produced near the 0.1 mbar level. In short, there is currently no consistent picture of Jupiter’s stratospheric circulations and how the distributions of trace species are impacted by those circulations.

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., 2003; Showman, 2003).
The observed temperature field features numerous wave signatures \cite{Li:2006, Fletcher:2017} and hotspots, both in the tropics and in the auroral regions \cite{Flasar:2004, Sinclair:2017}. Furthermore, a large variability of stratospheric temperature from one Earth year to another is observed at Jupiter’s equator \cite{Fletcher:2016}. This is associated with the quasi-quadiennal oscillation (QQO), a periodic oscillation in zonal wind and temperature thought to result from wave-mean zonal flow interactions \cite{Orton:1991, Flasar:2004, Simon-Miller:2007, Cosentino:2017}. The aforementioned stratospheric circulations are also putatively controlled by wave activity in the troposphere and stratosphere, although this remains to be characterized by quantitative studies capable to go beyond the simple analogy with the Brewer-Dobson circulation in the Earth’s stratosphere \cite{Butchart:2014}.

Understanding in detail the radiative forcings in Jupiter’s atmosphere is also key to interpreting the observed thermal structure. At mid-latitudes, a 5 to 10 K temperature contrast is reported between the summer and winter hemispheres despite Jupiter’s low obliquity ($3^\circ$) \cite{Fletcher:2016}. This has been supposedly attributed to radiative forcing due to aerosols, but has not been studied yet quantitatively. On a related topic, Zhang et al. \cite{Zhang:2015} recently highlighted the importance of radiative forcing by stratospheric aerosols of auroral origin, which were found to dominate the radiative heating at mid- and high- latitudes (instead of methane). Besides, focusing on the zonal-mean temperature in the upper troposphere (100–500 hPa), the cloudy equatorial zone is found to be 2–4 K colder than the warmer and clearer north and south equatorial belts, while there is little meridional temperature contrast at mid-latitudes (30$^\circ$N–70$^\circ$N and 30$^\circ$S–70$^\circ$S) \cite{Fletcher:2016}.

The aforementioned findings suggest that a complex interplay of dynamical and chemical activity takes place in Jupiter’s middle atmosphere, modulating the thermal structure and the distribution of trace species and aerosols, which in turn impact the radiative budget and dynamics. All these observations and open questions motivate the development of a new Global Circulation Model (GCM) for Jupiter extending to the upper stratosphere. Such a model would eventually take into account three-dimensional (3-D) dynamics, radiative forcings, photochemistry, cloud/aerosol microphysics and the couplings between them, including troposphere-stratosphere interactions. Several attempts have been made in this direction \cite{Yamazaki:2004, Medvedev:2013} illustrating the modeling complexity and high computational cost necessary to address the questions opened by observations. Recently, Young et al. \cite{Young:2019} developed a Jupiter GCM running at high resolution ($0.7^\circ$ in latitude $\times$ longitude), but their studies did not address stratospheric dynamics.

Our goal is to obtain a GCM capable to combine radiative transfer with high-resolution dynamics, akin to the approach we followed for Saturn’s atmosphere \cite{Guerlet:2014, Spiga:2018}. In this paper, we focus on the efficient parametrization of a radiative-convective model in Jupiter’s upper troposphere and stratosphere in the framework of an under-development GCM that aims at studying both tropospheric and stratospheric circulations. In this context, the calculation of heating and cooling rates has to be robust and fast, as these quantities need to be computed on a high-resolution grid (equivalent to about half a degree in latitude $\times$ longitude), every few Jupiter days and for several Jupiter years. These requirements exclude line-by-line models and require instead a broadband approach, as was recently done by Kuroda et al. \cite{Kuroda:2014} for Jupiter.
and by Guerlet et al. (2014) for Saturn’s atmosphere.

Apart from being part of a GCM, such a 1-D radiative-convective model can be used to compute radiative timescales (Kuroda et al., 2014) and, when confronted to observations, can be a useful tool to diagnose whether to first order radiative processes govern or not the thermal structure of the atmosphere (e.g. Guerlet et al., 2014). Kuroda et al. (2014) have developed such a radiative transfer model of Jupiter’s stratosphere using spectroscopic parameters from the HITRAN 2012 database and focusing on the validation of their correlated-k model. They took into account radiative forcing by the main gaseous species \( \text{H}_2, \text{He}, \text{CH}_4, \text{C}_2\text{H}_6 \) and \( \text{C}_2\text{H}_2 \) – but not ammonia), studied the impact of varying the hydrocarbon mixing ratio vertical profile on the thermal structure, but neglected the radiative impact of stratospheric aerosols, which are thought to have a major impact on the heating and cooling rates at mid and high latitudes (Zhang et al., 2015). In addition, although Kuroda et al. (2014) ’s model covered Jupiter’s upper troposphere, it accounted neither for tropospheric aerosols nor for clouds that are expected to play an important role in heat absorption and redistribution. In their paper, only comparisons with Galileo equatorial temperature profiles were presented, while many more observations of the meridional temperatures variations are available from Cassini and ground-based telescopes. In this paper, we propose to refine the approach proposed by Kuroda et al. (2014) by include the missing radiative contributions and to extend the comparisons of our seasonal model to the most recent observations.

Finally, knowledge of heating and cooling rates (diabatic forcings) can also be exploited to diagnose the residual-mean circulation in the stratosphere, as was done by West et al. (1992) and Moreno and Sedano (1997). This permits to estimate the stratospheric circulation and transport of tracers without building a GCM (under the limiting assumption that eddy heat flux is negligible compared to diabatic forcings). The studies by West et al. (1992) and Moreno and Sedano (1997) were based on observations from the Voyager epoch, and an update of this type of work based on Cassini-era observations and an up-to-date radiative transfer model would be valuable. This is especially critical because, while West et al. (1992) and Moreno and Sedano (1997) both agreed on the importance of including heating by stratospheric aerosols, the circulations they obtained differ both quantitatively and qualitatively.

In what follows, we describe a state-of-the art radiative-convective model for Jupiter’s atmosphere (as a necessary step for an under-development GCM), present comprehensive comparisons to recent observations and exploit the radiative heating field to compute the residual-mean circulation in Jupiter’s stratosphere. Section 2 describes our radiative transfer model for Jupiter’s upper troposphere and stratosphere that includes up-to-date spectroscopic parameters, an internal heat flux, radiative effects of tropospheric clouds and aerosols as well as stratospheric aerosols comprising fractale aggregates. We present the resulting thermal structure and compare it with recently published ground-based and Cassini observations in Section 3. We then detail our methodology to compute the residual-mean circulation in Section 4 and discuss these results in Section 5.
2 Jupiter radiative-convective model

2.1 Overall description

Our Jupiter radiative-convective model is adapted from its Saturn counterpart, described in detail in Guerlet et al. (2014). The two giant planets Jupiter and Saturn share many characteristics and, as a result, the main physical parametrizations are the same: a $k$-distribution model is used to 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 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 taken into account in our radiative 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 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 between Saturn and Jupiter radiative models relate to the gaseous composition as well as cloud and haze properties, detailed below.

2.2 Gaseous opacities and $k$-distribution model

Our Jupiter radiative model takes into account gaseous opacity from the three main hydrocarbons: methane (CH$_4$), ethane (C$_2$H$_6$), acetylene (C$_2$H$_2$), along with collision-induced transitions by H$_{2}$-H$_{2}$ and H$_{2}$-He. Through their infrared emissions, these molecules are the major stratospheric coolants, while atmospheric heating is primarily due to absorption of visible and near-infrared solar radiation by methane and aerosols. Furthermore, we also take into account opacity from ammonia (NH$_3$) that was previously neglected in the Saturn model, as is justified in section 2.4.

As line-by-line calculations of absorption coefficients are too time-consuming for the GCM runs we are aiming at, we use the correlated-$k$ method for the computation of the atmospheric transmission at each time step. Correlated-$k$ coefficients are pre-tabulated offline on a 2D temperature-pressure grid comprising twelve temperatures points from 70 to 400K and nine pressure levels from 10 to $10^{-6}$ bar (one level every pressure decade, plus one level at 0.5 bar as ammonia varies rapidly with altitude in this region). To obtain these tables, high-resolution absorption coefficient spectra $k(\nu)$ are first computed using the KSPECTRUM line-by-line model (Eymet et al., 2016).
for a mixture of gases (CH$_4$, C$_2$H$_6$, C$_2$H$_2$, NH$_3$) at each point of the (T,p) grid. The gaseous abundance profiles used are detailed in section 2.4. A Voigt line shape is assumed except for CH$_4$, for which we use the far-wing line shape of Hartmann et al. (2002), adapted to an H$_2$ atmosphere. In a second step, we discretize these spectra in large bands (typically 100–300 cm$^{-1}$ wide) and use the KDISTRIBUTION code (Eymet et al., 2016) to compute correlated-k coefficients $k(g)$ for each spectral band and each (T,p) value. We sample the cumulative probability $g$ with 8 Gauss integration points from 0 to 0.95 and another 8 points from 0.95 to 1. The spectral discretization and the number of bandwidths is a compromise between accuracy (which increases when small spectral intervals are chosen) and computation time. After multiple tests, we have selected 20 bands in the thermal infrared (10–3200 cm$^{-1}$ or 3 µm – 1 mm) and 25 bands in the visible and near-infrared (2000–33000 cm$^{-1}$ or 0.3–5 µm). When running a radiative-convective simulation, these $k$-coefficients are interpolated at each time step to the local temperature and to the pressure grid of the radiative transfer model. All radiative-convective simulations presented in this paper use a pressure grid consisting of 64 levels between 3 and 10$^{-6}$ bar.

2.3 Updates on methane spectroscopy

Spectroscopic line parameters are extracted from the HITRAN 2016 database (Gordon et al., 2017). However, the CH$_4$ linelist is known to be incomplete beyond 7,900 cm$^{-1}$; in particular, a methane absorption band at 1 µm is missing entirely, which could lead to an underestimation of the atmospheric heating rates. To fill this gap, we complete the HITRAN 2016 methane linelist with a recent linelist based on *ab initio* calculations (Rey et al., 2016, 2018). This list is available in the 0–12,000 cm$^{-1}$ range and contains position, energy and intensity for nearly 3.5 millions of transitions (assuming an intensity cut-off of 10$^{-28}$ cm/molecule), where the HITRAN 2016 database contains about 340,000 transitions. In order to limit the computation time, and because the HITRAN 2016 methane database is thought to be reliable up to 7,900 cm$^{-1}$, we choose to combine the two linelists, using the spectroscopic parameters of Rey et al. (2016) only beyond 7,900 cm$^{-1}$. Furthermore, we now include the transitions of the isotopologues $^{13}$CH$_4$ and CH$_3$D that were previously neglected by Guerlet et al. (2014). The isotopic ratio $^{13}$C/$^{12}$C is set to the terrestrial value (0.011) in agreement with Galileo measurements, and the ratio CH$_3$D/CH$_4$ to 7.79×10$^{-5}$ (Lellouch et al., 2001). The spectroscopic line data of $^{13}$CH$_4$ and CH$_3$D, for the spectral domain 0–12,000 cm$^{-1}$, also come from *ab initio* calculations by Rey et al. (2016), which are more exhaustive than HITRAN 2016.

Figure 1 shows the comparison between absorption coefficient spectra in the visible range computed using the HITRAN 2016 database (considering $^{12}$CH$_4$ only) with the new combination of the HITRAN 2016 and Rey et al. (2018) linelists for $^{12}$CH$_4$, $^{13}$CH$_4$ and CH$_3$D. This figure illustrates the important addition of the $^{12}$CH$_4$ Rey et al. (2018) linelist to HITRAN 2016 beyond 7,900 cm$^{-1}$, as well as the contribution of CH$_3$D that features emission bands at 1,100 cm$^{-1}$ (not shown), 2,200 cm$^{-1}$ and 3,500 cm$^{-1}$. $^{13}$CH$_4$ lines are not visible in this figure as their main absorption bands are mingled with $^{12}$CH$_4$. Regarding the impact on the equilibrium temperature profile, using the $^{12}$CH$_4$ linelist from Rey et al. (2018) beyond 7,900 cm$^{-1}$ increases the heating rates by 10% to 20% compared to using HITRAN 2016 alone in the range 0–12,000 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 \( \sim 1 \) 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 CH\(_4\) is concerned) up to 7,800 cm\(^{-1}\) with another set of \( k \)-distribution coefficients computed in the range 7,800–25,000 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 cm\(^{-1}\) was significant, but we did not distinguish between the near infrared part (7,800-12,000 cm\(^{-1}\)) and the visible part beyond 12,000 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 12,000–25,000 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 0.4K at the 10-20 mbar level. This can be explained by the small volume mixing ratio of methane combined to its rapidly decreasing absorption coefficients beyond 12,000 cm\(^{-1}\).

Hence, in this paper we choose to work only with the HITRAN 2016 and Rey et al. (2018) linelists and neglect gaseous absorption in the visible part (that is, beyond 12,000 cm\(^{-1}\)). The near infrared part of the spectrum is discretized in 22 spectral intervals, shown in Figure 1, to which we add three bands covering the visible part with zero gaseous opacity. These bands are needed, numerically, to contain cloud and aerosol opacity even in the absence of gaseous opacity.

### 2.4 Gaseous abundances

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

The volume mixing ratio of CH\(_4\) decreases with altitude primarily due to molecular diffusion in the upper stratosphere, and to a lesser extent due to photo-dissociation by solar UV radiation (Gladstone et al., 1996; Moses et al., 2005). The altitude level of the methane homopause on Jupiter is estimated to lie in the range \( 10^{-5} \) to \( 10^{-6} \) bar based on stellar occultations (Festou et al., 1981; Yelle et al., 1996; Greathouse et al., 2010). This is significantly deeper than on Saturn, where the homopause level is estimated to a few \( 10^{-7} \) bar. This difference is explained by a much stronger eddy mixing coefficient on Saturn compared to Jupiter (Moses et al., 2005). However, the exact homopause level on Jupiter is not well constrained by observations and differs among studies (Greathouse et al., 2010); it could also vary with time and latitude. Similarly, uncertainties on the eddy mixing coefficient and photodissociation rates map into a family of the methane vertical abundance profile in Jupiter photochemical models (see for instance models A, B and C of Moses et al. (2005)).

The choice of a vertical profile for methane (and other hydrocarbons) is thus partly arbitrary and will influence the vertical energy deposition, hence the resulting equilibrium temperature profile.
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) or also including Rey et al. (2018) linelists for $^{12}\text{CH}_4$ (beyond 7,900 cm$^{-1}$) and for CH$_3$D and $^{13}\text{CH}_4$ (in the range 0–12,000 cm$^{-1}$), in blue. The vertical dashed lines represent the limits of the 22 bands used for generating $k$-distribution coefficients.
in the stratosphere, as was already reported by Kuroda et al. (2014). We choose to set the volume mixing ratio of the three hydrocarbons to the average of the 1-D photochemical models A and C from Moses et al. (2005). This corresponds to an homopause level at \( \sim 1 \mu \text{bar} \). Regarding \( \text{C}_2\text{H}_6 \) and \( \text{C}_2\text{H}_2 \), we further scale these model profiles so that the hydrocarbon abundances at 1 mbar match the low to mid-latitude Cassini/CIRS observations of Nixon 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 sensitivity tests, we also compiled a different set of \( k \)-tables with the hydrocarbon profiles set to the photochemical model profiles used by Nixon et al. (2007), which feature a deeper homopause level (\( \sim 10 \mu \text{bar} \)). The different hydrocarbon vertical profiles are illustrated in Fig. 2.

We present in Fig. 3 the impact of assuming different hydrocarbon profiles on the equilibrium temperature, based on aerosol-free 1-D radiative-convection simulation for latitude 20°N. In the 1 mbar to 10 \( \mu \text{bar} \) region, the Nixon et al. (2007) photochemical model has \( \sim 1.5 \) to 3 times more acetylene than our combination of the Moses et al. (2005) models (but similar amounts of ethane and methane), resulting in greater cooling rates and stratospheric temperatures 2 to 5K colder. Between 1 and 10 \( \mu \text{bar} \), 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 \( \mu \text{bar} \)) 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 \( \mu \text{bar} \).

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 \( \pm 8 \)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 actually divided by two: this 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 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 compensation of the two competing effects (an increase in radiative cooling rates when ethane is increased, a decrease of it when acetylene is decreased). 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 cooler 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.
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°S and 40°N (shown as squares).
Figure 3: Equilibrium temperature profiles for different hydrocarbon mixing ratio profiles. Left: example at latitude 20°N, Ls= 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 C$_2$H$_2$ and C$_2$H$_6$ (in red). Right: example at latitude 60°N, Ls= 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 C$_2$H$_6$ and a 50% decrease in C$_2$H$_2$ (red line).
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 $\mu$m range.

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$150K) (Atreya et al., 1999). Following the vapour pressure curve, its mixing ratio rapidly decreases above this pressure level to become insignificant at tropopause levels. We set the ammonia "deep" mixing ratio (at 3 bar) to 250 ppm consistently with planet-average abundances measured by Juno at this pressure level (Li et al., 2017) and assume a fractional scale height of 0.15 above the 0.7 bar level (Nixon et al., 2007). We find that including NH$_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 NH$_3$ and also by a small greenhouse effect. Indeed, adding ammonia increases the infrared opacity, especially beyond 5 $\mu$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 (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 10–20°N (as low as 140 ppm at 1 bar) ([Li et al., 2017]). The influence of these spatial variations on the thermal structure is deferred to a future study. This new finding for Jupiter does not challenge our previously published results on Saturn’s thermal structure ([Guerlet et al., 2014]). Indeed, the upper tropospheric volume mixing ratio of NH$_3$ is 10 to 100 times lower on Saturn than on Jupiter, due to a deeper condensation level (~1.4 bar), and we find that including 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 thrived 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 (0.3–2 μm) located above a compact cloud comprising larger particles (3–100 μm) is needed ([e.g. West et al. (1986) from Pioneer data, Irwin et al. (2001) using Galileo/NIMS spectra, Wong et al. (2004a) using Cassini/CIRS data]).

The location of the cloud deck is estimated to lie in the range 0.5–1.2 bar depending on the studies, while the upper tropospheric haze likely extends up to 150–300 mbar, i.e. near the tropopause. Thermochemical equilibrium models predict that ammonia condenses at ~700 mbar, while ammonium hydrosulfide (NH$_4$SH) is expected to form another cloud layer at ~2 bar ([Atreya et al. 1999]). However, the spectroscopic signatures of NH$_3$ ice at 2 μm, 9.4 μm and 26 μm have been rarely observed, and [Baines et al. (2002)] showed that spectrally identifiable ammonia clouds cover less than 1% of Jupiter’s globe. Rather, [Sromovsky and Fry (2010)] suggest that the haze layer consists of small ammonia-coated particles overlying a cloud layer of NH$_4$SH ice particles at ~600 mbar, or that several layers of NH$_3$ and NH$_4$SH ice particles coexist, which would explain the lack of strong NH$_3$ absorption features in the infrared. A similar conclusion was reached by [Giles et al. (2015)], who used Cassini/VIMS data between 4.5 and 5.2 μm to constrain Jupiter’s cloud structure. The authors find that VIMS observations can be modeled using a compact, highly reflecting cloud layer located at a pressure of 1.2 bar or lower, with spectrally flat optical properties in this spectral range. Indeed, setting the refractive index to that of pure NH$_3$ or NH$_4$SH ice particles could not fit VIMS observations, for any particle sizes in the range 1–40 μm.
A few observational constraints exist on haze and cloud particles optical properties: Pioneer observations analyzed by Tomasko et al. (1978) require highly reflecting particles at 0.44 and 0.6 \( \mu m \), 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 \( \mu m \)) 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 opacities of the order of \( \sim 1 \) have been derived at 2 \( \mu m \) (Irwin et al., 2001). The opacity variations between cloudy zones and less opaque belts likely stem from cloud opacity variations (found to lie between 8 and 22 at 5 \( \mu m \)) 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 opacity and physical properties at a global scale. 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), opacity, 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 NH\(_3\) ice particles, with optical constants from Martonchik et al. (1984);
2. pure NH\(_4\)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));
4. particles with nearly grey optical constants.

Composition 4 has real and imaginary indices set arbitrarily close to that of NH\(_3\) except for smoother spectral features (since the sharp absorption features of 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 NH\(_4\)SH particles (composition 2) are expected to be brighter than the other kind of particles, as a result of the low real index of NH\(_4\)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 \( \mu m \).

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

Overall, we find that all cases considering a pure NH$_4$SH cloud lead to a too bright albedo (>0.55), regardless of the assumptions on cloud opacity, cloud particle size or haze properties. Similarly, all cases with “Saturn”-like haze particles combined with ammonia cloud particles result in too dark albedos (∼0.4), which is consistent with the high imaginary index of these haze particles in the visible. A combination of a dark “Saturn” haze with a bright NH$_4$SH cloud leading to a ∼0.5 albedo might be found, but we choose to discard solutions with the “Saturn” haze as several studies (e.g., Tomasko et al. 1978) suggest that Jupiter’s haze particles must have a larger single scattering albedo than Saturn’s. The “grey” and NH$_3$ haze particles considered here are in better agreement with estimates of the single scattering albedo by Tomasko et al. (1978).

Albedos comparable to that reported by Li et al. (2018b) are obtained for combinations of “grey” and/or NH$_3$ particles for the haze and cloud material, with the condition that the haze opacity amounts to 3–4. Different combinations of the nature of the haze and cloud particles (NH$_3$–NH$_3$, grey–grey, grey–NH$_3$) give similar results, which is not surprising given their similar optical constants. We confirm that small haze particles (∼0.5 µm) are needed in order to reproduce the 3 to 4 times larger haze optical depth observed in the visible compared to the near infrared: with the haze optical depth set to 4 at 0.75 µm for ammonia or grey particles, the opacity amounts to ∼1 at a wavelength of 2 µm, which is compatible with observations by Irwin et al. (2001).

The heat deposition differs depending on the cloud and haze composition, integrated opacity and altitude of the cloud deck, as illustrated in Fig 6. For scenarios with bright NH$_4$SH clouds, the heating rate increases 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 opacity 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 opacity as well as the altitude of the cloud deck on the temperature. In our “grey haze, NH$_3$ cloud” scenario, increasing the cloud opacity 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 to 660 mbar results in warming the troposphere by 5 K (while having little impact on the albedo, of the order of 2%), as more solar light is absorbed at higher altitudes. Because the resulting temperature at the 1-bar level is closer to observations when setting the cloud deck at 840 mbar, we choose to keep this setting throughout
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 type | Cloud Size (in µm) | τ cloud at 750 nm | Haze type | Haze Size (in µm) | τ haze at 750 nm | Bond Albedo |
|------------|--------------------|-------------------|-----------|-------------------|-----------------|-------------|
| NH₄SH      | 15.00              | 7.00              | Grey      | 0.50              | 2.00            | 0.59        |
| NH₄SH      | 15.00              | 7.00              | Grey      | 1.00              | 2.00            | 0.59        |
| NH₄SH      | 15.00              | 7.00              | Grey      | 0.50              | 4.00            | 0.61        |
| NH₄SH      | 15.00              | 10.00             | Grey      | 0.50              | 2.00            | 0.63        |
| NH₄SH      | 15.00              | 7.00              | NH₃       | 1.00              | 2.00            | 0.59        |
| NH₄SH      | 15.00              | 7.00              | “Saturn”  | 1.00              | 2.00            | 0.55        |
| NH₄SH      | 15.00              | 15.00             | “Saturn”  | 1.00              | 2.00            | 0.62        |
| NH₃        | 15.00              | 7.00              | “Saturn”  | 1.00              | 2.00            | 0.39        |
| NH₃        | 15.00              | 7.00              | “Saturn”  | 1.00              | 4.00            | 0.43        |
| NH₃        | 15.00              | 15.00             | “Saturn”  | 1.00              | 2.00            | 0.40        |
| NH₃        | 15.00              | 15.00             | “Saturn”  | 1.00              | 4.00            | 0.43        |
| NH₃        | 15.00              | 4.00              | Grey      | 0.50              | 2.00            | 0.41        |
| NH₃        | 15.00              | 10.00             | Grey      | 0.50              | 2.00            | 0.43        |
| NH₃        | 15.00              | 7.00              | Grey      | 1.00              | 2.00            | 0.42        |
| NH₃        | 15.00              | 7.00              | Grey      | 2.00              | 2.00            | 0.42        |
| NH₃        | 10.00              | 7.00              | Grey      | 0.50              | 2.00            | 0.44        |
| NH₃        | 20.00              | 7.00              | Grey      | 0.50              | 2.00            | 0.41        |
| NH₃        | 15.00              | 7.00              | Grey      | 0.50              | 4.00            | 0.48        |
| NH₃        | 10.00              | 10.00             | Grey      | 0.50              | 4.00            | 0.50        |
| NH₃        | 10.00              | 15.00             | Grey      | 0.50              | 4.00            | 0.51        |
| NH₃        | 10.00              | 10.00             | NH₃       | 0.50              | 4.00            | 0.51        |
| Grey       | 10.00              | 10.00             | Grey      | 0.50              | 4.00            | 0.48        |
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. In this example, the haze and cloud optical depth at 0.75 μm are set to 4 and 10, respectively, and the haze and cloud particle sizes to 0.5 and 10 μm. The cloud deck is set at 840 mbar except for one case with a slightly shallower cloud deck, in blue.
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 opacity 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 retroactions with meridional circulation – not yet included – are expected to play an important role. 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 µm on top of a NH₃ cloud with 10-µm particles, a cloud deck at 840 mbar with an integrated opacity at 750 nm of 15.

2.6 Stratospheric aerosols

2.6.1 Observational constraints and motivation

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

- One that has the same composition and particle size as the tropospheric haze layer but is optically thin, similarly to what is done in Guerlet et al. (2014). We set its integrated optical depth to 0.1.

- One that is not uniform with latitude and that is more absorbant in the UV, described further below.

The addition of this second kind of aerosol is motivated by the observations of dark polar hoods at near-UV wavelength (Hord et al., 1979; Tomasko et al., 1986), which have been attributed to a stratospheric haze layer. This haze is both forward scattering and strongly polarizing, which implies that it is composed of aggregate particles similar to Titan’s haze particles (West and Smith, 1991). The favored scenario for their formation is through precipitation 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. (2013) 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 Subsysteme (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 ~20 mbar at high latitudes (60–70°). Regarding the aerosol sizes and shapes, ISS observations can be fitted with small sub-micron spherical particles at low latitudes. Poleward of 30°N and 45°S, ISS observations are consistent with fractal aggregates composed of a thousand monomers of 10-nm radius for a fractal dimension of 2, corresponding to an effective radius of about 0.7µm.
Zhang et al. (2013) also derive 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).

Based on these observational constraints, Zhang et al. (2015) show that this haze dominates the radiative heating budget at middle and high latitudes in Jupiter’s middle stratosphere, with a contribution of the haze reaching up to 10 times the heating rate due to CH\textsubscript{4} alone in the 10–20 mbar pressure range. Hence, radiative heating by the polar haze appears to be a key component to be included in our radiative-convective equilibrium model, which parametrization is detailed below.

### 2.6.2 Parametrization of the aerosol properties

The optical properties of fractal aggregates haze particles (extinction coefficient, scattering albedo and asymmetry factor) are computed using a semi-empirical model from Botet et al. (1997). This model employs the mean-field approximation in the case of scattering of an electromagnetic wave by a cluster of monosized spheres. We compute the optical properties for aggregates with a fractal dimension of 2, monomers of 10 nm, and a log-normal distribution for the number of monomers centered at 1000. The imaginary index $n_i$ is set to 0.02 at 255 nm and 0.001 at 900 nm as constrained by Zhang et al. (2013) from Cassini/ISS observations, with a logarithmic interpolation in between as in Zhang et al. (2015). The real index is set to 1.65 in the visible range, similar to the mean value of Zhang et al. (2015). In the thermal infrared, due to the lack of observational constraints, we adopt the real and imaginary index of Guerlet et al. (2015) derived from Cassini/CIRS observations of Saturn’s auroral aerosols, which are thought to be of similar origin and nature.

Regarding the meridional variations of the polar haze opacity, we build a meridional profile based on that retrieved by Zhang et al. (2013), with a slightly smoother transition at mid-latitudes where the opacity varies by several orders of magnitude over a few degrees of latitude. The integrated haze opacity as a function of latitude adopted in our model is compared with Zhang et al. (2013) retrievals in Figure 7. Poleward of 70° (the highest latitude observed in Zhang et al., 2013), we simply assume that the haze opacity is equal to that at 70°.

We choose to parameterize the aerosol opacity vertical profile with the following function:

$$\tau(p) \propto (p/p_m)^2 \times \exp \left(-\left(p/p_m\right)^2\right)$$

(1)

with $\tau$ the optical depth, $p$ the pressure, and $p_m$ parametrizing the pressure level of the haze layer. This function reproduces well, to first order, the retrieved vertical profile of the haze opacity by Zhang et al. (2013). 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°, as derived by Zhang et al. (2013). In our model, the opacity profile is then normalized at each latitude bin so that the opacity integrated above the 80 mbar level matches the meridional profile in Fig. 7. The resulting vertical profiles of the haze opacity are shown at three latitudes in Figure 7 along with those retrieved by Zhang et al. (2013). We note that our parameterized profiles decrease less steeply with altitude than that observed. A Gaussian-shaped profile actually provides a better match to the observed profiles and is tested as well, but results in local heating rate and temperature maxima that are too sharp and do not match the observed temperature profiles (this point is further detailed in section 3). Furthermore,
the slower decrease in opacity with altitude in our parametrized profiles, is in better agreement with predictions from microphysical-photochemical models (Friedson et al., 2002).

We note that Zhang et al. (2013) retrievals suggest that, poleward of 60°, the tropospheric aerosol layer shifts to lower pressure levels (~100 mbar instead of 200 mbar), which we did not take into account (our tropospheric layer extends up to 180 mbar at all latitudes, with a thin stratospheric haze layer on top of it).

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°N and 50 and 70°S. Figure 8 shows an example at 60°S, Ls= 180°, where the heating rate is increased by a factor of 6 at the 10-mbar pressure level when stratospheric aerosols are included, which is in 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 temperature profile. At this latitude and season, we find that it warms the atmosphere by up to 20K at the 10 to 20-mbar pressure level (see Figure 8). This effect decreases with pressure, amounting to 3–4K at the 1-mbar pressure level.

At high latitudes (> 75°) the net radiative impact of the polar haze depends on season: during winter, the net effect 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.
Figure 8: Daily-averaged heating rate, in Kelvin per Earth day (left) and temperature (right) vertical profiles at latitude 60°S and Ls= 0° including (in red) or not (in black) the polar haze contribution. For this latitude and season, the absolute value of the cooling rate nearly lies on top of the heating rate.
3 Thermal structure and comparisons to observations

3.1 Internal heat flux and tropospheric temperature

Before presenting in detail the results of our radiative-convective model at equilibrium, we address the issue of the tropospheric equator-to-pole temperature gradient. Indeed, it is well known since the Pioneer 10 and 11 era that Jupiter exhibits no significant temperature (or emitted thermal infrared flux) latitudinal gradient at the 1-bar level (Ingersoll, 1976). However, assuming a uniform internal heat flux $F_{\text{cst}} = 7.48 \text{ W.m}^{-2}$, our radiative-convective model produces a strong temperature contrast of 28K at 1 bar between the warmer equator (178K) and colder poles (150K) (see Figure 12). This is expected from such a radiative model, as solar insolation is maximum at low latitudes all year round (given Jupiter’s low obliquity).

To explain the observed near-uniform tropospheric temperatures, several theories have been proposed. For instance, using a turbulent, 3-D deep convection model, Aurnou et al. (2008) finds that convective heat transfer by quasi-geostrophic thermal plumes results in an outward heat flow 2.5 to 3.2 times greater at the poles than at equator. This latitudinal trend is consistent with the work of Pirraglia (1984) who tried to estimate the meridional variations of internal heat flux needed to reconcile the observed solar energy deposition with the outgoing thermal radiation. On a different note, with their General Circulation Model for Jupiter’s troposphere, Young et al. (2019) found that even when considering a uniform internal heat flux, atmospheric dynamics acts to balance the latitudinal-varying solar forcing. As a consequence, the 1-bar equator-to-pole temperature gradient is reduced from 35K with a radiative-convective version of their GCM to only 5K when using their full GCM with resolved atmospheric dynamics.

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

$$F_{\text{int}}(\theta) = 0.67 \times F_{\text{cst}} + 0.66 \times F_{\text{cst}} \times \sin(\theta) \times \sin(\theta)$$  \hspace{1cm} (2)

$$F_{\text{int}}(\theta) = 0.5 \times F_{\text{cst}} + F_{\text{cst}} \times \sin(\theta) \times \sin(\theta)$$  \hspace{1cm} (3)

This ad hoc parametrization ensures a planet-average internal heat flux equal to $F_{\text{cst}}$ while setting an internal heat flux twice larger (eq. 2) or three times larger (eq. 3) at the poles than at the equator. When using eq. 2, the equatorial temperature at the 1-bar level is now 9K warmer than the poles (instead of 28K when a uniform internal heat flux is assumed). The associated outgoing thermal emission is only 8% larger at the equator than at 60° latitude, which is consistent with Pirraglia (1984) observations, which extended to 60° only. However, when using eq. 3, the temperature is actually 2K warmer (and the outgoing thermal emission 8% larger) at 60° than at the equator. Hence, in what follows, we discuss the thermal structure obtained with eq. 2 which yields more realistic results. It is worth emphasizing here that the temperature field at pressures lower than ~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. The corre-
Figure 9: Pressure-latitude cross-section of the temperature (in K) in Jupiter’s atmosphere at Ls=0°. The internal heat flux varies with latitude as defined in eq. 2.

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 (i.e. overall 5 K colder than our model predictions). At latitudes 50°–70°, 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 as reported in section 2.6. Finally, at latitudes poleward of 70°, temperatures are the coldest with a 100K tropopause and a maximum stratospheric temperature of 140K to 150K at the 1-mbar level. We note that at high latitudes, the tropopause is broader and extends from 100 mbar to 20 mbar, which is caused by the heating by CH₄ being less efficient in the lower stratosphere due to the low solar elevation.

Seasonal variations are expected to be small owing to Jupiter’s low obliquity. We present in
Figure 10 shows the seasonal evolution of the 10-mbar temperature at 50°N and 50°S. We first note that the amplitude of seasonal variations is very small in the southern hemisphere: it is only 2 K at 50°S, whether or not the radiative impact of auroral aerosols is taken into account. This can be explained by the competing effects of obliquity and eccentricity, as Jupiter’s perihelion occurs at Ls=57° close to southern winter. On the other hand, these two effects add up in the northern hemisphere, where the peak-to-peak seasonal amplitude is ~6 K for the auroral aerosol-free case. When auroral aerosols are included, there is a global temperature increase of 12 K in the northern hemisphere and 7 K in the southern hemisphere. This difference can be explained by the hemispheric asymmetry in aerosol content as constrained by Zhang et al. (2013), where the integrated opacity is about four times greater at 50°N compared to 50°S (see Figure 7). The peak-to-peak amplitude of seasonal variations is also slightly enhanced at 50°N (8 K instead of 6 K) when we include this additional aerosol radiative forcing. The seasonal amplitude reported here for 50°N and 10 mbar is typical of the northern stratosphere (latitudes 45°N – 75°N) between 30 mbar and 0.01 mbar. Finally, at 50°N, when auroral aerosols are added, we also notice that the temperature maximum is shifted to an earlier season (Ls=100° instead of Ls=125°), closer to northern summer solstice, which hints at shorter radiative timescales as a result of adding auroral aerosols (see next section 3.3 for further details).

3.3 Radiative timescales

In this section, we evaluate and discuss radiative relaxation timescales in Jupiter’s atmosphere. Radiative timescales can be used to assess whether the atmosphere responds quickly or not to changes in atmospheric temperatures and solar insolation. It is sometimes used in 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 detailed inventory of the radiative forcings, as is done in this study. Recent estimates by Kuroda et al. (2014) and Li et al. (2018a), based on their respective radiative-convective equilibrium models, take into account gaseous radiative forcings similar to ours, but neglect all kind of clouds and aerosols.

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

Two examples are shown in Figure 11 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 (~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
Figure 10: 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 50°N and 50°S, as labeled. Two cases are shown, including or not auroral aerosols.
Figure 11: Two example profiles of the radiative timescale in Jupiter’s atmosphere at $L_s=0$ at the equator (in blue) or at 40°N where auroral aerosols are abundant (in red). These are compared to values published in Kuroda et al. (2014) for the equator (stars).

The 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 auroral aerosol radiative forcing 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 all aerosols in their model. All the conclusions in this paragraph hold when we compare our results to the recent work by Li et al. (2018a).

### 3.4 Comparison to observations

The monitoring of Jupiter’s spatio-temporal variations in temperature from the analysis of thermal infrared spectra started with the Voyager spacecrafts in 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). Nowadays, these studies are pursued using the Texas Echelon Cross Echelle Spectrograph (TEXES) instrument on the IRTF (Lacy et al., 2002; Fletcher et al., 2016; Sinclair 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 2–4° in latitude [Fletcher et al., 2016], which is actually comparable to the spatial resolution achieved by the Cassini fly-by. Both CIRS and TEXES are sensitive to the temperature in the pressure range 700–0.5 mbar (with the caveat of a low sensitivity in the 20–60 mbar range) and cover the latitude range 78°S – 78°N.

In this section, we focus on the comparison with the results of [Fletcher et al., 2016] who analysed spectra acquired by TEXES in December, 2014 (corresponding to Ls=175°) and also analysed, with the same retrieval pipeline, observations from the Composite Infrared Spectrometer (CIRS) on board Cassini during the December, 2010 flyby (corresponding to Ls=110°). In our comparisons, we neglect longitudinal variability and only consider temperatures derived from zonally-averaged spectra provided by [Fletcher et al., 2016].

We first focus on the comparison in the upper troposphere, shown in Figure 12. At the 360-mbar level, the temperature derived from CIRS and TEXES shows little meridional or temporal variability except at the equator, where the TEXES-derived temperature is about 5K warmer in 2014 than the CIRS-derived temperature in 2000. These variations are attributed to changes in the dynamics of the equatorial belts [Fletcher et al., 2016]. As already mentioned in section 2.5.2, the cloudy equatorial zone is colder than warmer, less cloudy, equatorial belts at 15°N and 15°S, which is thought to be the consequence of vertical motions (upwelling in zones, subsidence in belts) rather than due to a radiative effect. Near the tropopause level (at 110 mbar), observed temperatures exhibit a small (5K) decrease in temperature from 50 to 78° in both hemispheres, and a temporal variability of the order of 3 K in the southern hemisphere.

Our predicted temperatures, obtained with the crude parametrization of a latitudinal-varying internal heat flux (as is defined in eq. 2) and a single cloud and haze scenario, reproduces reasonably well the observed globally-averaged temperature in the troposphere. Our main disagreement is that at the 360-mbar level, our temperatures are ~4K cooler than observations in the 40–78° latitude range. The fact that our model still slightly underestimates the temperature at high latitudes suggests that we should let the internal heat flux increase even more with latitude. However, because this is only a crude parametrization that might become obsolete when full GCM simulations are run, we did not attempt to tune eq. 2 until we obtained a perfect match with observations. In addition, other processes could be at play in this model-observation mismatch: indeed, the aerosol opacity cross-section derived by [Zhang et al., 2013] indicate that the tropospheric haze layer could extend at higher altitudes in the 60–70° latitude range (with no observations beyond 70°), which could enhance the radiative heating in the upper troposphere. Given the current lack of observational constraints on the haze properties at high latitudes, we did not modify our tropospheric haze scenario.

Regarding the stratosphere, we present in Figure 13 the meridional temperature variations at four pressure levels: 0.4, 3, 10 and 25 mbar. The comparison between our radiative-equilibrium temperatures and CIRS and TEXES data is also displayed at four latitudes, in the form of temperature vertical profiles, in Figure 14. 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-quadrennial oscillation (QOO) in the temperature and associated thermal wind field thought to result from wave-mean zonal flow interactions [Orton et al., 1991].
Figure 12: 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, $L_s=110$).
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 an excellent agreement with observations. At this level, both CIRS and TEXES feature a local temperature maximum of 160 K at 50–65°N which is well reproduced by our model, should the auroral aerosols be taken into account. If not, the temperature would be 10 K to 15 K colder at this pressure level and latitude range. The hemispherical asymmetry between latitudes 60°N and 60°S, of about 8 K, observed by TEXES and CIRS, is faithfully reproduced as well. We can determine that it is caused by a radiative effect related to the “auroral” haze absorption, as this hemispherical asymmetry would be of only 1–2 K without this radiative contribution. As already discussed in section 3.2 and shown in Figure 10, this asymmetry is primarily due to the much higher amount of haze opacity in the northern hemisphere compared to the southern hemisphere, and not so much to a seasonal effect (temperatures at 50–60°N are expected to be warmer than that at 50–60°S throughout the year, if the haze hemispheric asymmetry persists). Moreover, in their analysis of Voyager observations, Simon-Miller et al. (2006) also found that the temperature was ~6 K warmer at 50°N compared to 50°S, which is compliant with our predictions. At the time of the Voyager encounter, the season was shortly after autumn equinox (Ls=190°).

Our model predicts that a similar north-south asymmetry of 10 K between 60°N and 60°S is still present at the 25-mbar level. This is at odds with CIRS and TEXES observations, which are nearly symmetric about the equator. This suggests that we overestimate the aerosol content in the northern hemisphere at this pressure level but not in the southern hemisphere, as our predicted temperature agrees well with CIRS and TEXES observations at 50–70°S. In other words, the actual vertical profile of aerosol opacity is probably more complex than the one parameterized in eq. 1 and might be quite different between 60°N and 60°S. However, we also note that CIRS and TEXES measurements have a rather low sensitivity to the temperature in the 20 to 60 mbar range, so that it is possible that part of the observation-model mismatch at 25 mbar is also due to a larger uncertainty in the observations at this level.

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

At lower pressure levels (p< 3 mbar), we note that our predicted temperatures are almost systematically underestimated by ~5 K compared to TEXES and CIRS observations. This suggests that either a radiative ingredient is missing in our model, or that the temperature is governed by other processes, such as dynamical heating by gravity wave breaking or by eddies. We note that at 0.4 mbar, TEXES and CIRS observations exhibit important temporal variability at high latitudes (poleward of 55°). The observed temperature also increases between 60 and 78° latitude in both hemispheres, which is at odds with our simple radiative-convective model. Clearly, other processes
Figure 13: 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, in solid lines (in black for $L_s=110$, in red for $L_s=176$). The four panels display temperatures at four different pressure levels (0.4, 3, 10 and 25 mbar). For reference, we also show the simulated temperature obtained when auroral aerosols are neglected (dashed line, $L_s=110$).
Figure 14: 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). For reference, we also show the temperature predicted by our model without auroral aerosols (dashed black line).

must control the temperature at these altitudes. One hypothesis is that the temperature may be influenced by the precipitation of high-energy particles that could warm the atmosphere at high latitudes through Joule heating (Sinclair et al., 2017).

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

This comparison work with state-of-the-art observations shows that our simple radiative-convective equilibrium model do reproduce well, to first order, the observed temperature in the upper troposphere and lower stratosphere (p>5 mbar), except in the equatorial region (where there is well-known dynamical activity). Other processes seem to be at play in controlling the temperature in the middle and upper stratosphere, and in the troposphere (belt/zone activity). In the following section, we take advantage of the computed heating and cooling rates to estimate the residual-mean circulation in Jupiter’s stratosphere.

4 Residual-mean circulation

4.1 Background

Stratospheric circulations are driven by a combination of diabatic and mechanical (eddy-induced) forcings, resulting in an interplay of transport processes: advection, stirring and mixing. On the Earth stratosphere, it has been shown that the Transformed Eulerian Mean (or residual-mean) circulation is a good approximation to the Lagrangian mean circulation (relevant to tracer transport) in regions where wave breaking and dissipation is relatively weak [Dunkerton, 1978; Butchart, 2014]. As we describe in what follows, the residual-mean circulation can be approximately estimated from the knowledge of atmospheric net heating rates and temperatures. Hereafter we follow this approach to diagnose the zonally-averaged mass circulation in Jupiter’s stratosphere, for annually-averaged conditions.

The complete equations for the zonally-averaged stratospheric circulation are provided by the Transformed Eulerian-mean formulation, with the respectively residual-mean meridional and vertical components of the circulation \((v^*, w^*)\) defined as a combination of a zonal-mean and eddy-induced terms

\[
v^* = \overline{v} - \frac{1}{\rho_0} \frac{\partial}{\partial z} \left( \rho_0 \frac{\overline{v'\theta'}}{\theta'} \right)
\]

\[
w^* = \overline{w} + \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \left( \cos \phi \frac{\overline{v'\theta'}}{\theta'} \right)
\]

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

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

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

\[
\frac{1}{a \cos \phi} \frac{\partial (\cos \phi v^*)}{\partial \phi} + \frac{1}{\rho_0} \frac{\partial (\rho_0 w^*)}{\partial z} = 0
\]
and an energy-conservation (thermodynamic) equation

\[
\frac{\partial \bar{\theta}}{\partial t} + \frac{v^*}{a} \frac{\partial \bar{\theta}}{\partial \phi} + w^* \frac{\partial \bar{\theta}}{\partial z} = Q + E
\]  

(8)

in which \( Q \) is the net radiative heating rate and \( E \) is the heating rate related to eddies forcing the mean flow

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

(9)

For quasi-geostrophic large-scale flows in non-acceleration conditions, the eddy-related term \( E \) 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 can also be neglected, which entails

\[
\frac{v^*}{a} \frac{\partial \bar{\theta}}{\partial \phi} + w^* \frac{\partial \bar{\theta}}{\partial z} \approx Q
\]  

(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 \( E \approx 0 \) (we neglect the eddy heat flux convergence term) and \( \bar{\theta}/\partial t \approx 0 \) (we consider annually-averaged conditions). In equation 10 we use the net radiative heating rates \( Q \) and temperature profiles \( \bar{\theta}(z) \) computed by our radiative equilibrium model for Jupiter. In other words, in this methodology, we estimate what would be the residual-mean circulation for a modelled planet undergoing the radiative forcings described in section 2 and exhibiting the atmospheric temperature field presented in section 3. We discuss the extent to which this residual-mean circulation can relate to actual conditions on Jupiter in section 4.3.

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)

\[
\Psi_{i=1} = \int_{-\frac{\pi}{2}}^{\phi} (w_{i=1}^* + \epsilon) \cos \phi a d\phi
\]  

(11)
where
\[
\epsilon = \left( \int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} \cos \phi \, d\phi \right)^{-1} \int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} w_{i=1}^* \cos \phi \, d\phi
\]  
(12)

is a corrective term (usually a couple percent at best) designed to ensure that the stream-function \( \Psi \) vanishes at the north pole.

3. The meridional component \( v_{i=1}^* \) is obtained from the streamfunction \( \Psi_{i=1} \) by using equation (vertical derivative of \( \Psi_{i=1} \)). Using the streamfunction \( \Psi_{i=1} \) to compute \( v_{i=1}^* \) from \( w_{i=1}^* \) is equivalent to solving the mass-conserving equation.

4. The meridional component \( v_{i=1}^* \) is injected in equation (10) to obtain the vertical component \( w_{i=2}^* \) at the next iteration, then the process is looped back to step 2 for \( i = 2 \).

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

### 4.3 Results and comparison to previous studies

The resulting streamlines of the residual circulation, for annual-averaged conditions and for the stratosphere only, are displayed in Figure 15 and the corresponding vertical and meridional wind speeds are shown in Figure 16. For reference, the radiative heating rates \( Q \) used to derive this circulation is shown in Figure 17. In the lower stratosphere (1 to 20 mbar), the residual-mean circulation is characterized by two meridional cells with rising motion at high latitudes, where the polar stratospheric haze results in significant net radiative heating, and subsidence at 15°S. The southern cell is confined to latitudes 15°S to 80°S while the northern cell extends from 80°N to 15°S. This north-south asymmetry stems from the asymmetry in the aerosol’s meridional distribution (and their associated radiative forcings). This asymmetry is clear in Figure 17: the region of net radiative heating is broader in latitude in the northern hemisphere, corresponding to a broader upwelling region.

In the lower and upper stratosphere, the residual-mean circulation exhibits upwelling motion at the equator and subsidence at both poles, which is a consequence of the net radiative heating at low latitudes (which receive greater solar insolation) and net radiative cooling at the poles. This is the expected trend in the absence of “auroral” stratospheric haze.

Previous estimates of the residual-mean circulation in Jupiter’s atmosphere were obtained by West et al. (1992) and Moreno and Sedano (1997) using the same underlying formalism as ours but with a slightly different approach. The authors used equations (7) and (10) under the same approximation (neglecting the eddy heat flux convergence term) but using the temperature field derived by Voyager/IRIS to compute the cooling rates. Instead, here we used the radiative-equilibrium temperature and heating/cooling rates computed on each grid point by our model.

On the one hand, to study the upper troposphere and tropopause region, using the observed temperature is undoubtedly more insightful. Indeed, the troposphere features horizontal temperature gradients at a 20° latitude scale, which are in thermal wind equilibrium with the alternately
Figure 15: Streamlines computed using eq. 11. The altitude is computed with the convention $z=0$ km at the 1-bar level. For reference, the bottom of the figure, at 70 km, corresponds to the lower stratosphere ($\sim 20$ 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).

Figure 16: Pressure-latitude cross-section of the vertical (left) and meridional (right) components of the residual-mean circulation, in m.s$^{-1}$. 
Figure 17: Pressure-latitude cross-section of the net heating rates in Jupiter’s stratosphere, in K.s$^{-1}$, for yearly-averaged conditions.
eastward and westward zonal jets. When solving equations 7 and 10, those meridional temperature gradients result in meridional gradients of the cooling rates and induce a circulation closely linked to the zonal jets (upwelling in zones, subsidence in belts, West et al., 1992). Our calculations cannot capture this effect by nature. This is the reason why we do not discuss our residual circulation obtained for pressures higher than ∼50 mbar.

On the other hand, our approach is more self-consistent in the stratosphere. Indeed, our temperature and cooling rates are obtained from a radiative-equilibrium model, and we also assume that radiative forcings dominate in the right-hand side of equation 8. Conversely, the observed temperature could be affected by eddy forcing terms. This was discussed in section 3.4, where we noted that the radiative-equilibrium modeled temperatures are ∼5 K cooler than the observed temperatures at all latitudes for pressures < 3 mbar and displayed significant latitudinal and temporal variability. Energy transfer by eddies could be important at these pressure levels. If that is the case, then neglecting $E$ in equation 8 is not valid. Hence, using the observed (warmer) temperatures for computing the cooling rates (as was done by West et al. (1992) and Moreno and Sedano (1997)) while neglecting $E$ when estimating the streamfunction could lead to erroneous results.

Furthermore, both West et al. (1992) and Moreno and Sedano (1997) encountered issues with the net radiative heating rate values computed from Voyager-IRIS temperatures. These temperature profiles had a low vertical resolution in the stratosphere and were provided only at 77 mbar and 1 mbar: for intermediate pressure levels, temperature was obtained by interpolation. West et al. (1992) notices that using this temperature field results in an imbalance between cooling and heating rates at global scale, with residual net radiative heating at the 10-mbar pressure level. Given the knowledge gained since those two studies were published, we propose that this issue was related to the latitudinally-varying radiative forcing of the “auroral” aerosols: while West et al. (1992) and Moreno and Sedano (1997) did take into account latitudinal-varying solar radiative heating by a polar haze, the Voyager/IRIS temperature field did not capture the localized temperature increase due to aerosols at high latitudes in the 10–30 mbar region (later seen by TEXES and Cassini/CIRS, see Figure 13), due to a poor vertical resolution. These warmer regions imply a locally larger cooling rate; ignoring this effect would explain the excess of net radiative heating reported by West et al. (1992) at the 10-mbar pressure level at global scale. To mitigate this issue, West et al. (1992) scaled either the heating rate, or the infrared cooling rate, by a factor that depends on altitude only, while actually their cooling rate was missing a latitudinal-varying term.

Despite those differences, the residual mean stratospheric circulation we obtain from our computations resembles closely that obtained by West et al. (1992) between 30 and 1 mbar. As can be seen in Figure 16, maximum vertical windspeed values are the order of 0.3 mm/s (upwelling) at 75° latitude and 3 mbar, with maximum equatorward motion of the order of 0.25 m/s at 3 mbar. This is twice as strong as the values derived by West et al. (1992) (actually shown in Friedson et al. (2002)), but is qualitatively consistent.

It would be interesting to evaluate the impact of this circulation on the distribution of ethane and acetylene, the main by-products of methane photochemistry. In their study, Hue et al. (2018) combine a photochemical model with a simple parametrization of transport (both advection and diffusion) with the goal of explaining the observed increase in ethane towards high latitudes, while acetylene is decreasing. In doing so, they test several Hadley-like circulation cells with upwelling at the equator and subsidence at both poles. Our study suggests that such a circulation is not
appropriate in the lower stratosphere, where these puzzling hydrocarbon distributions are observed. It might be possible, however, that changing the circulation regime in the lower stratosphere would not help reconciling models and observations: an equatorward transport of hydrocarbons would indeed deplete even more the high latitudes in ethane.

Regarding CO\(_2\) and HCN, products of the SL-9 impact at 44°S, one way to reconcile their opposite trends (CO\(_2\) being maximum at the south pole while HCN is found well-mixed from mid-southern to mid-northern latitudes) would be if CO\(_2\) was deposited at pressures either lower than 0.1 mbar (190 km altitude in Figure 15) or around 10 mbar (90 km) to be transported poleward, while HCN was deposited near 0.5–1 mbar, to be transported equatorward. As already noted by Lellouch et al. (2006), this represents a significant difference in altitude that is difficult to explain. Another lead is that chemical processes are at play to increase CO\(_2\) at high southern latitudes, maybe linked with heterogeneous processes on stratospheric auroral aerosols. Further work on this topic would help understanding their distributions.

5 Discussion and Conclusions

We have developed a radiative-convective equilibrium model for Jupiter’s troposphere and stratosphere that includes parametrizations of several cloud and haze layers. As for its Saturn counterpart (Guerlet et al., 2014), this model is computationally efficient and aims at being coupled with a dynamical core of a General Circulation Model (as was recently done in the Saturn case, see Spiga et al., 2018). We take into account the radiative contribution of:

- CH\(_4\), C\(_2\)H\(_6\), C\(_2\)H\(_2\) and NH\(_3\) for radiatively active species along with collision-induced absorption from H\(_2\)-H\(_2\), H\(_2\)-He;
- A rather compact ammonia cloud located at 840 mbar comprising 10-\(\mu\)m particles, with a visible integrated opacity of 10.
- A tropospheric haze layer extending between 660 and 180 mbar composed of 0.5 \(\mu\)m particles with near ”grey” optical constants and an integrated opacity of 4.
- A stratospheric haze layer made of fractal aggregates (1000 monomers of 10-nm each, fractal dimension of 2) supposedly linked with precipitation of high-energy particles. Their opacity is maximum near the 20-mbar level and at latitudes poleward of 30°N and 45°S.

The gaseous abundance profiles as well as tropospheric cloud and haze layer properties are fixed in latitude and time in the current version of our model, but we have studied the sensitivity to varying those parameters. For instance, varying the abundance of hydrocarbons with latitude in similar proportions than the observed ones (i.e. a poleward enhancement in ethane 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 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 stratospheric “aerosols” have an important role in the radiative budget of Jupiter’s stratosphere. Their net impact at latitudes 45–60° is to warm the atmosphere in the 5–30 mbar range, by up to 20 K. 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. The hemispheric asymmetry in stratospheric aerosols opacity (that is much larger in the northern than in the southern hemisphere) combined with the small obliquity of Jupiter causes 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 (Fletcher et al., 2016), observed at the time of the Cassini flyby in 2000 (Ls=110°) or during a TEXES observing run in 2014 (Ls=176°), is well reproduced by our model. Nevertheless, it remains uncertain whether the observation by Zhang et al. (2013) (the adopted scenario in our model) of a larger amount of haze in the northern hemisphere compared to the southern hemisphere is a permanent feature, or if the aerosol field undergoes strong temporal variability.

Our results regarding the very good comparison of our radiative-convective model with observations at and around the 10-mbar level suggests that the temperature might be, to first order, controlled by radiative processes at this pressure level (except for equatorial latitudes, where the QQQ signature dominates). This is consistent with the relatively short radiative timescales estimated in section 3.3 (100 days, or 3% of a Jupiter year).

At lower pressures (p<3 mbar), 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. Given that the hydrocarbon abundances are well documented at these low pressures, that we set their abundance to a model profile in agreement with the aforementioned observations, and that we use state-of-the-art spectroscopic linelists, it is unlikely that this mismatch comes from a wrong evaluation of the gaseous heating rates. It is also unlikely that a missing stratospheric haze, not accounted for in our model, is responsible for this warming at all latitudes, as this haze would have already been observed otherwise. We thus favor a mechanical forcing such as gravity wave breaking or other eddy terms warming the atmosphere. A coupling with thermospheric or ionospheric circulations is also a possibility, through Joule heating, adiabatic compression or horizontal advection (e.g. Majeed et al., 2005, although the upper stratosphere is marginally covered).

The knowledge of net radiative heating rates can be then exploited to estimate the stratospheric residual-mean circulation, under the assumption that the eddy heat flux convergence term is negligible, in a similar fashion as West et al. (1992). In the Earth’s stratosphere, the residual-mean circulation represents well, to first order, the transport of tracers in regions where wave breaking and dissipation are weak (Butchart, 2014, and references therein). This topic is of high interest on Jupiter, as the observed meridional distribution of photochemical products (ethane and acetylene) or by-products of comet Shoemaker Levy 9 impact (HCN, CO₂, dust...) is puzzling and cannot be explained by simple chemistry-transport models. In the lower stratosphere (1–30 mbar), we find that the residual-mean circulation is characterized by upwelling at high latitudes (due to
net radiative heating by the polar stratospheric haze) and subsidence at 10°S. These results are qualitatively consistent with a previous study by West et al. (1992). In the middle and upper stratosphere, the circulation turns into Hadley-like circulation cells with rising motion at the equator and subsidence at both poles, consistently with the net radiative heating at low latitudes and cooling at high latitudes.

It would be interesting to evaluate the impact of such a circulation regime on the distribution of trace species, including the by-products of comet SL-9 (CO₂, HCN...) but also hydrocarbons. Explaining the distributions of ethane and acetylene in Jupiter’s stratosphere remains an open question, and it is worth mentioning that a similar dichotomy in ethane and acetylene is observed in Saturn’s lower stratosphere (Guerlet et al., 2009; Sinclair et al., 2013) and remains an open question as well. Drawing a comparative study of Jupiter and Saturn on that topic would probably be fruitful, especially given the different seasonal forcings on those two giant planets. Finally, Friedson et al. (1999) have shown that the residual-mean circulation obtained by West et al. (1992), very similar to ours, was too weak to account for the spreading of SL9 dust debris in the upper troposphere and lower stratosphere but that including eddy-forcing terms provided a better match to the observations. Future work should then focus on better estimating these terms.

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 what is the residual-mean circulation associated with the computed 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 – 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 driving the QQO. These topics are also valid for Saturn’s atmosphere, opening the way to comparative studies between these two gas giants.

Acknowledgements

S. Guerlet and A. Spiga acknowledge funding by the French Agence Nationale de la Recherche (ANR) under grant agreements ANR-12-PDOC-0013 and ANR-14-CE23-0010-01. Part of this work was funded by CNES as a support for Cassini/CIRS data interpretation. We thank Xi Zhang (UC Santa Cruz) for providing their retrieved aerosol number density map, Jeremy Burgalat and Pascal Rannou (Reims University) for sharing their library to generate optical constants for fractal aggregates and Michael Rey (Reims University) for providing linelists for methane and its isotopologues (available through the TheoReTS plateform: http://theorets.univ-reims.fr/).

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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