Greenhouse effect: the relative contributions of emission height and total absorption

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ABSTRACT

Since the 1970’s, results from radiative transfer models unambiguously show that an increase in the CO$_2$ concentration leads to an increase of the greenhouse effect. However, this robust result is often misunderstood and often questioned. A common argument is that the CO$_2$ greenhouse effect is saturated (i.e. does not increase) as CO$_2$ absorption of an entire atmospheric column, named absorptivity, is saturated. This argument is erroneous firstly because absorptivity by CO$_2$ is currently not fully saturated and still increases with CO$_2$ concentration, and secondly because a change in emission height explains why the greenhouse effect may increase even if the absorptivity is saturated. However, these explanations are only qualitative. In this article, we first propose a way of quantifying the effects of both the emission height and absorptivity and we illustrate which one of the two dominates for a suite of simple idealized atmospheres. Then, using a line by line model and a representative standard atmospheric profile, we show that the increase of the greenhouse effect due to an increase of CO$_2$ from its current value is primarily due (about 90%) to the change in emission height. For an increase of water vapor, the change in absorptivity plays a more important role (about 40%) but the change in emission height still has the largest contribution (about 60%).

1. Introduction

To establish the physical laws that govern the surface temperature of a planet, Fourier (1824; 1837) made the analogy between a vessel covered with plates of glass and the Earth surface covered by the atmosphere (Pierrehumbert, 2004). Using this framework, Arrhenius (1896) made the first estimate of the greenhouse effect and of the sensitivity of the surface temperature to a change in CO$_2$ concentration of the atmosphere. His computation was based on a single layer model where the surface was covered by an isothermal atmosphere for which the outgoing long-wave flux at the top-of-the-atmosphere (TOA) reads:

\[
\hat{F} = \frac{T_s}{\tau} B(T_s) + \left( 1 - \frac{T_s}{\tau} \right) B(T_a)
\]

where $B(T)$ is the black body emission, i.e. the Stefan-Boltzmann law, for a temperature $T$, $\tau$ is the total broadband hemispherical transmissivity, i.e. the transmissivity for radiation crossing the whole atmosphere, from its top to the surface, averaged over the longwave domain (overline variables refer to variables averaged over the longwave domain) and over an hemisphere. As we assume scattering in the longwave domain is negligible, the broadband absorptivity of the atmosphere in the longwave domain is equal to $1 - \frac{T_s}{\tau}$ and is equal to the broadband emissivity of the atmosphere. $T_s$ is the surface temperatures and $T_a$ a bulk temperature of the atmosphere, generally called emission temperature. The broadband greenhouse effect, de-

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The “saturation paradox” can be summarized as follows: why does the greenhouse effect increase with the CO₂ concentration (Fig. 1-b) whereas the broadband absorptivity does not increase as much, especially when water vapor is present (Fig. 1-a)? As highlighted by Eq. 2, the absorptivity is not the only main parameter that controls the greenhouse effect; the emission temperature $T_a$ of the atmosphere is also fundamental. If the increase of CO₂ concentration has little impact on absorptivity, it has a significant impact on $T_a$. When the CO₂ increases, the infrared radiation that escapes toward space is emitted by the atmosphere at a higher altitude. As most of the radiation is emitted by the troposphere, higher altitude means lower emission temperature, lower value of the Planck function, lower value of the radiation emitted toward space and therefore higher value of the greenhouse effect (Hansen et al., 1981; Pierrehumbert, 2010; Archer, 2011; Benestad, 2017). For a doubling of the CO₂ concentration, the average value of the change in emission height is about 150 m, assuming that the radiative forcing of about $\approx 4\text{ W m}^{-2}$ can be translated into a change in black body temperature emission, and then into a change in emission height assuming a temperature vertical gradient of $\approx 6.5\text{ K km}^{-1}$ (Held and Soden, 2000).

Beyond the single layer model, for fundamental physical reasons, the increase of the greenhouse effect due to an increase of the concentration of an absorbing gas, in particular CO₂, is partly due to an increase of absorptivity and partly due to an increase of emission height (Pierrehumbert, 2010). However, the contribution of each of these two effects has not been quantified yet, and the main goals of this paper are to present a framework that allows quantifying the contribution of these two effects, and to perform the quantification. A second goal is to quantify the change in emission height, and not only its impact on the flux at the TOA. This offers the possibility to propose a new quantitative simplified description of the greenhouse effect that is more realistic than the too simple single layer model called blanket model (Benestad, 2017).

In this study, we will use only prescribed atmospheric profiles and will therefore compute the forcing when changing the absorbing gas concentration. All calculations are for cloudless skies. In section 2 we present the framework that allows to separate and quantify the contribution of absorptivity and that of emission height to the flux at the tropopause, and therefore to the greenhouse effect. To allow some analytical developments, especially for simple limiting cases, we consider monochromatic radiances and idealized vertical atmospheric profiles. In section 3 we still consider radiances but with realistic atmospheric profiles. This will help us to interpret the results presented in section 4, where we compute the flux at the tropopause over the whole thermal infrared domain and where we independently increase the concentration of the two most important greenhouse gases on Earth, H₂O and
CO₂. The temperature adjustment of the stratosphere is also analyzed. Finally, summary and conclusion are given in section 5.

2. Formulation with simplified conditions

To present the main concepts and to facilitate some analytical developments, we first consider the very simple case of an idealized atmosphere where the monochromatic absorption coefficient is constant along the vertical and the volumetric mass density only depends on pressure and therefore on altitude \(z\)

\[ \rho(z) = \rho(0) e^{-z/h_t} \]  

(3)

where \(h_t\) is the scale height (\(h_t \approx 8\,\text{km on Earth}\)). Hydrostatic pressure follows a law of the same type as density: \(P(z) = P(0) e^{-z/h_t}\). To quantify the radiative forcing of CO₂, it has been shown that the change of the net flux at the tropopause is much more relevant than the change of the net flux at the top of the atmosphere (Shine et al., 1995; Hansen et al., 1997; Stuber et al., 2001). To keep the atmospheric profile as simple as possible, we ignore the stratosphere in a first step and then show Sect. 4-d that this simplification has little impact for the key points addressed in this study. We therefore consider the troposphere only, i.e. an atmosphere, which vertical extent ends at the tropopause. A last simplification is to assume that the temperature vertical profile is such that the monochromatic radiance emitted by a black body (or Planck function) \(\mathcal{B}(P)\) increases linearly with pressure \(P\) (Fig. 2):

\[ \mathcal{B}(P) = \mathcal{B}(P_t) + \frac{P - P_t}{P_t - P_s} (\mathcal{B}(P_t) - \mathcal{B}(P_s)) \]  

(4)

where \(P_t = 1000\,\text{hPa}\) and \(P_s = 200\,\text{hPa}\) are the pressure at the surface and at the tropopause, \(H_s = 0\) and \(H_t = h_t \log(P_t/P_s) \approx 12.9\,\text{km}\) are the altitude of the surface and the tropopause, respectively. Note that curly letters refer to monochromatic directional variables. We consider two contrasted profiles (Fig. 2): a profile where \(\mathcal{B}\) decreases from the Planck function at surface \(\mathcal{B}_s\) to a value \(\mathcal{B}_t\) at the tropopause \((\mathcal{B}(P_t) = \mathcal{B}_t\) and \(\mathcal{B}(P_s) = \mathcal{B}_m\)) and an isothermal profile chosen so that the two profiles have the same mass weighted mean value: \(\mathcal{B}(P_t) = \mathcal{B}(P_t) = \mathcal{B}_m = \mathcal{B}_t + \mathcal{B}_s\). The Planck function is computed for a wave number \(\nu_c = 550\,\text{cm}^{-1}\) (corresponding to a wavelength \(\lambda_c \approx 18\,\mu\text{m}\) close to the strong CO₂ 15\,\mu\text{m} absorption band and for temperatures \(T_s = 294\,\text{K}\) \((\mathcal{B}_s \approx 0.144\,\text{Wm}^{-2}\text{sr}^{-1})\) and \(T_t = 220\,\text{K}\) \((\mathcal{B}_t \approx 0.056\,\text{Wm}^{-2}\text{sr}^{-1})\). We assume the atmosphere has a homogeneous concentration of absorbing gases and we neglect the effects of pressure and temperature on the specific absorption coefficient \(k\) (in \(\text{m}^2\text{kg}^{-1}\)). Therefore \(k\) is constant along the vertical. The surface is assumed to be a perfect black body. We also assume that radiation propagates only along the vertical, which allows to replace the integral on the zenith angle by considering one single angle. The radiative exchanges are computed at a given frequency and with a formalism adapted for general plane parallel atmospheres (Schwarzkopf and Fels, 1991).

a. Basic equations and the limiting case of the single layer model

With the above assumptions, the expression of the optical thickness between the tropopause and a layer of pressure \(P\) at altitude \(z\) simplifies as

\[ \tau(P) = k f (P - P_t)/g \]  

(5)

where \(g\) is the gravity in \(\text{m}\,\text{s}^{-2}\). We introduce \(f\), which is a multiplicative factor to allow a proportional change in absorption within the whole atmosphere. By default, \(f = 1\). The spectral outgoing radiance \(\mathcal{F}\) at the tropopause in the zenith direction then reads (Pierrehumbert, 2010):

\[ \mathcal{F} = \mathcal{F}_s \mathcal{B}_s + \int_{P_t}^{P_s} \frac{\partial \mathcal{F}(P)}{\partial P} \mathcal{B}(P) dP \]  

(6)

where \(\mathcal{F}(P)\) is the directional transmissivity between altitude of pressure \(P\) and the tropopause

\[ \mathcal{F}(P) = e^{-\tau(P)} \]  

(7)

\(\mathcal{F}_s = \mathcal{F}(P_s) = e^{-\tau_s}\) is the transmissivity of the troposphere with \(\tau_s = \tau(P_s)\) the total optical thickness of the troposphere, i.e. from the tropopause to the surface. \(\mathcal{F}_s = \mathcal{F}_t \mathcal{B}_s\) is the radiance that is emitted by the surface and that

\[ \mathcal{F}_t = \mathcal{F}(P_t) = e^{-\tau_t}\]
reaches the tropopause. \( \mathcal{F}_d \) is the vertical integral of the radiance that is emitted by the troposphere and that reaches the tropopause.

Equation 6 may be written as:

\[
\mathcal{F} = \mathcal{F}_s B_e + (1 - \mathcal{F}_s) F_e
\]  

(8a)

where \( B_e \) is the equivalent blackbody emission of the atmosphere:

\[
B_e = \int_P^{P_t} B(P) \omega(P) dP
\]  

(8b)

and

\[
\omega(P) = \frac{1}{1 - \mathcal{F}_e} \frac{\partial \mathcal{F}(P)}{\partial P}.
\]  

(8c)

Equation 8a has the same form as the classical single layer model (Eq. 1) except that here it is spectrally resolved. This equation makes explicit that the radiance at the tropopause depends directly on the transmissivity \( \mathcal{F}_s \), and therefore on the total optical thickness \( \tau_s \) of the troposphere. This transmissivity \( \mathcal{F}_s \) impacts both the radiance that reaches the tropopause emitted by the surface and the radiance that reaches the tropopause emitted by the troposphere.

The equivalent blackbody emission \( B_e \) of the atmosphere is the mean value of the blackbody emission \( B(P) \) weighted by the \( \omega(P) \) function (Eq. 8b). For optically very thin atmospheres (\( \tau_s \ll 1 \)), \( B_e \) is equal to the pressure-weighted mean of \( B(P) \). Then \( B_e \approx [B(P_t) + B(P_s)]/2 \) and is the same for the two idealized atmospheric profiles considered in this section (Appendix Sect. A1).

When the troposphere is optically thin, the radiance \( \mathcal{F} \) at the tropopause decreases when the optical thickness \( \tau_s \) of the troposphere increases, starting from a value \( \mathcal{F} = B_s \) when \( \tau_s = 0 \) (Fig. 3, black line). As long as \( \tau_s \ll 1 \), the decrease of the radiance \( \mathcal{F} \) at the tropopause is proportional to \( \tau_s \) and similar for both atmospheric profiles because the pressure-weighted mean temperature of both troposphere is the same (see Sect. A1-a in Appendix). When the troposphere is isothermal, this decrease gradually slows down from optical thickness \( \tau_s \) larger than 0.5 and reaches a plateau when the optical thickness is larger than about 4. When the troposphere is non-isothermal, the slowdown is not as fast as for the isothermal case and the decrease continues for optical thickness larger than 4. The limiting value of the radiance at the tropopause for infinite value of the optical thickness is much smaller (and therefore the greenhouse effect much higher) in the non-isothermal case compared to the isothermal case.

The model generally used in simplified explanations of the greenhouse effect (Eq. 1) assumes that the troposphere is isothermal along the vertical. With this assumption, the flux at the tropopause does not decrease any more when the total optical thickness \( \tau_s \) increases if \( \tau_s \) is larger than 4. It is then said that the greenhouse effect “saturates”. This saturation effect almost disappears when the temperature decreases with height: The greenhouse effect continues to increase when the optical thickness \( \tau_s \) increases, even for large value of \( \tau_s \). For a non-isothermal troposphere the altitude where the emitted radiation escapes to space matters. We now present how this effect of emission height can be quantified.

b. Contribution of absorptivity and emission height to radiance changes

The sensitivity of the radiance \( \mathcal{F} \) at the tropopause to a fractional change in amount of absorbing gases reads, according to Eq. 8:

\[
\frac{\partial \mathcal{F}}{\partial f} \equiv \mathcal{F}' = \frac{\partial \mathcal{F}_s}{\partial f} (B_s - B_e) + (1 - \mathcal{F}_s) \frac{\partial B_e}{\partial f}
\]  

(9)

which we rewrite as:

\[
\mathcal{F}' = \mathcal{F}'_s + \mathcal{F}'_e
\]  

(10a)

\[
\mathcal{F}'_s = \frac{\partial \mathcal{F}_s}{\partial f} (B_s - B_e)
\]  

(10b)

\[
\mathcal{F}'_e = (1 - \mathcal{F}_s) \frac{\partial B_e}{\partial f}
\]  

(10c)

These three terms are shown in Fig. 4 as a function of the total optical thickness \( \tau_s \) of the troposphere.

According to Eq. 10b, \( \mathcal{F}'_s \) quantifies how much the radiance \( \mathcal{F} \) at the tropopause is directly impacted by a change
in the transmissivity $T_a$, and therefore by a change in the absorptivity $A_s = 1 - T_a$, when the amount of absorbing gases changes. $\mathcal{F}_\tau$ is the sensitivity of the radiance at the tropopause if the troposphere is isothermal, or would be isothermal, at a temperature that corresponds to a black body emission $B(T)$ (i.e. $\partial B/\partial f = 0$). For both temperature profiles, the absolute value of $\mathcal{F}_\tau$ linearly increases with $\tau_a$, is maximum for $\tau_a \approx 1$, becomes very small for $\tau_a$ larger than 4 and is almost zero when the troposphere is fully opaque (Fig. 4). $\mathcal{F}_\tau$ is slightly higher for the non-isothermal profile as $B(T)$ has a smallest value with this profile compared to the isothermal profile (Fig. 3).

$\mathcal{F}_{Z_e}$ quantifies how much the radiance at the tropopause is impacted by a change in $Z_e$ when the amount of absorbing gases changes. Radiance $B_e$ (Eq. 8b) is the weighted average of the Planck function over the whole troposphere with a weight $\omega(P)$ (Eq. 8c), which depends on the optical exchange factor between the atmosphere at pressure $P$ and the tropopause (Dufresne et al., 2005). This weight varies from a function that is constant with pressure when the total optical thickness is low ($\tau_a \ll 1$) to a function that is maximum at the tropopause, decreases with increasing pressure and is almost zero close to the surface when the total optical thickness is large ($\tau_a \gg 1$) (Fig. 5, and Sect. A1-b in Appendix). As a consequence, the radiation that reaches the tropopause is emitted on average at lower pressure, i.e. at higher altitude, when the optical thickness of the troposphere increases. It is said that the “emission height” increases (Hansen et al., 1981; Held and Soden, 2000; Pierrehumbert, 2010; Archer, 2011; Benestad, 2017). The variable $\mathcal{F}_{Z_e}$ quantifies how much this change in emission height impacts the radiance at the tropopause. It is zero for an isothermal troposphere since $\partial B_e/\partial f = 0$. If the temperature of the troposphere decreases with height, an increase of emission height yields a decrease of the temperature, a decrease of the Planck function and therefore a decrease of the upward radiance at the tropopause. The sensitivity $\mathcal{F}_{Z_e}$ due to change in emission height increases with the total optical thickness $\tau_a$ of the troposphere, reaches a maximum for $\tau_a \approx 4$, and then slowly decreases (red line on Fig. 4).

c. Emission height

After defining the contribution of the change in emission height to the change in radiance at the tropopause, we now define the emission height itself. Since we assume in this section that the absorption coefficient $k$ is constant, the optical thickness increases linearly with pressure (Eq. 5). Therefore, many radiative variables are easier to compute and to interpret in pressure coordinate rather than in altitude coordinate. We will therefore continue to write the equations in pressure coordinate, and the “emission height” will be defined as the altitude corresponding to the “emission pressure”.

The relative contribution $\Omega(P)$ of a layer of thickness $dP$ at pressure $P$ to the radiance $\mathcal{F}_a$ reads, according to Eq. 6:
When the troposphere is isothermal, $\mathcal{B}(P) = \mathcal{B}_e$ and therefore $\Omega(P) = \omega(P)$. The probability density that a photon emitted by the troposphere and that reached the tropopause has been emitted at a level of pressure $P$ is equal to the probability density that a photon going downward at the tropopause is absorbed at level of pressure $P$. This is consistent with the reciprocity principle. Therefore, when the total optical thickness $\tau_e$ is small, the emission pressure is equal to the average between the tropopause pressure and the pressure at surface, i.e. $(P_t + P_s)/2$. The corresponding altitude is slightly higher than 4km, which is consistent with what is observed in Fig. 6. Compared to the isothermal profile, $Z_e$ is lower when the temperature decreases with height as the Planck function gives more weight to the lower and warmer part of the troposphere.

Starting from the altitude where the pressure is $(P_t + P_s)/2$, the emission height $Z_e$ increases when the total optical thickness $\tau_e$ increases (Fig. 6). The emission height $Z_e$ is commonly approximated as the altitude where the optical thickness is one (Pierrehumbert, 2010; Huang and Bani Shahabadi, 2014). For the profiles considered here, this approximation is valid as soon as the total optical thickness is larger than about 4 (Fig. 6).

3. Results with more realistic cloudless atmospheres

We now abandoned previous idealized vertical profile and consider a realistic cloudless atmosphere, namely
and therefore the vertical profile of the Planck function at $\nu_c = 550\text{cm}^{-1}$ is not linear with pressure anymore. The volumetric mass density varies according to the perfect gas law and the atmosphere is discretized into 65 vertical layers. The CO$_2$ concentration is 287 ppmv as in Collins et al. (2006). We perform the same computations as in the previous section with this new profile, and the results show little differences compared to those displayed on Figures 3 to 6 (not shown). The exact values are slightly modified but all the key features are identical.

### a. Single absorption line

We now consider a narrow frequency range around the center of an absorption line instead of a given frequency. Molecules in gases have discrete energy levels and absorption of photons correspond to transitions between these discrete energy levels. The absorption lines are very numerous (many millions) and not infinitely sharp due to broadening mechanisms. In the Earth troposphere, pressure broadening (also named collision broadening) is the dominant effect and will be the only one considered in a first step. In the vicinity of a line center, the spectral absorption coefficient $k$ varies with frequency according to a Lorentzian profile:

$$k = \frac{S}{\pi} \frac{\alpha}{(\nu - \nu_c)^2 + \alpha^2}$$  \hspace{1cm} (15)

where $S$ is the line absorption integrated intensity, $\nu$ is the wavenumber, $\nu_c$ the wavenumber of the line center and $\alpha$ is the half-width at half-height. Lorentz half width is assumed be proportional to $PT^{-0.5}$

$$\alpha(P,T) = \alpha_0 \frac{P}{P_0} \left(\frac{T_0}{T}\right)^{0.5}$$  \hspace{1cm} (16)

with $\alpha_0 \approx 0.1\text{cm}^{-1}$ at $P_0 = 1013\text{hPa}$ and $T_0 = 300\text{K}$, which are typical values for the CO$_2$ lines around 550cm$^{-1}$. We assume that the line intensity is constant along the vertical and is multiplied by a factor $f = 1$ to allow a proportional change in absorption within the whole atmosphere, as in previous section.

The sensitivity of the spectral radiance at the tropopause to a fractional change in the absorbing gas for a line can be deduced from single frequency results (Fig. 4). A first example is shown for a single and weak absorption line (Fig. 7). The optical thickness at the absorption line center is about 0.75 and decreases rapidly away from the line center. The sensitivity of the radiance is maximum at the line center and is primarily due to the change in absorptivity ($F'_T$, blue line) as the optical thickness is small.

This picture is very different for a line whose absorption intensity is 12 times larger and that will be referred later as a line of “medium intensity” (Fig. 8). Around the absorption line center, the sensitivity $F'_T$ due to a change in absorptivity is zero as one may expect from Fig. 4. In

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The mid latitude summer (MLS) atmospheric profile (McCleachey et al., 1972; Anderson et al., 1986) that has often been used to benchmark radiative codes (Ellingson et al., 1991; Collins et al., 2006; Pincus et al., 2015).

As in the previous section, we consider an atmosphere that ends at the tropopause. The temperature and pressure at the surface and at the tropopause are close to the previous idealized profile ($P_s = 1013\text{hPa}$, $H_s = 0\text{m}$, $T_s = 294\text{K}$, $P_t = 190\text{hPa}$, $H_t = 12.6\text{km}$, $T_t = 218.4\text{K}$). The vertical profile of temperature is almost linear with altitude
this spectral region the sensitivity \( S' \) due to a change in emission height is the dominant factor. Close to the absorption line center, the optical thickness is large and the mean emission height is located close to the tropopause (Fig. 9). An increase of optical thickness has little impact on the emission height. Away from the absorption line center, the sensitivity due to a change in emission height is still the dominant factor and an increase of absorbing gas decreases the radiance at the tropopause and therefore increases the greenhouse effect. Further away from the absorption line center, the sensitivity due to a change in total absorption dominates, and slowly decreases away from the absorption line center.

### 4. Radiative flux over the whole infrared domain with realistic radiative and thermodynamic properties

Until now we considered only radiances that allowed us to avoid angular integration. In this section we show how the framework based on radiances can be easily transposed to a framework for irradiance, or radiative flux. We use the classical approximations for pristine atmospheres. The atmosphere is absorbing and non-scattering, perfectly stratified along the horizontal (plane parallel assumption) and the surface has an emissivity of one.

#### a. Framework for radiative flux

With the above assumptions, the spectral flux \( F \) at the tropopause is (Pierrehumbert, 2010; Dufresne et al., 2005):

\[
F = \tilde{F} B_s + \int_{P_t}^{P_i} \frac{\partial \tilde{F}(P)}{\partial P} B(P) dP
\]

where \( B = \pi B \) and \( \tilde{F}(P) \) is the spectral hemispherical transmissivity between the pressure level \( P \) and the pressure at the tropopause level \( P_t \):

\[
\tilde{F}(P) = 2 \int_0^1 \exp(-\tau(P, \mu)) \mu d\mu
\]

where \( \tau(P, \mu) \) is the spectral directional optical thickness between the tropopause and the pressure level \( P \)

\[
\tau(P, \mu) = \left| \int_{P_t}^{P} f \frac{k(P)}{8 \mu} dP \right|
\]

where \( \mu \) is the cosine of the zenith angle, \( k(P) \) the specific absorption coefficient at level of pressure \( P \), and \( f = 1 \) a multiplicative factor as in the previous sections. Equation 17 is similar to Eq. 6 previously used, except that the radiances \( \tilde{F}(\beta, B) \) have been replaced by the irradiance (or flux) \( F, B \), and the directional transmissivity \( \tilde{F} \) has been replaced by the hemispherical transmissivity \( \tilde{F} \). With these replacements, one can show that equations 8 to 14 can be directly adapted to fluxes. For instance, Eq. 17 can be rewritten as:

\[
F = \tilde{F}_B + (1 - \tilde{F}_s) B_e
\]

\[
B_e = \int_{P_t}^{P_i} B(P) \omega(P) dP
\]

\[
\omega(P) = \frac{1}{1 - \tilde{F}_s} \frac{\partial \tilde{F}(P)}{\partial P}
\]

which can be compared to Eqs. 8a-8c. In Eq. 20a, \( \tilde{F}_B \) is termed the surface transmitted irradiance in Costa and Shine (2012). Applying the same replacements to Eqs. 10a-10c allows to split the sensitivity of the flux at the tropopause \( F' \) in a contribution \( F'_{s,i} \) due to the change in absorptivity and a contribution \( F'_{e,i} \) due to the change in emission height.

Many radiative codes do not compute the sensitivity \( F' \) directly, and a difference in radiances can therefore be more suitable. For two atmospheres \( i = 1, 2 \) that only differ by the amount of absorbing gases, the flux at the tropopause reads:

\[
F_i = \tilde{F}_{s,i} B_s + (1 - \tilde{F}_{s,i}) B_{e,i}
\]
One can show (see Sect. A2 in appendix) that the
difference between the two fluxes \( \Delta F = F_2 - F_1 \) reads as:

\[
\begin{align*}
\Delta F &= \Delta F_\text{f} + \Delta F_\text{ze} \\
\Delta F_\text{f} &= (\tilde{T}_{\text{s},2} - \tilde{T}_{\text{s},1}) [B_3 - B_{e,1}] \\
\Delta F_\text{ze} &= (1 - \tilde{T}_{\text{s},2}) [B_{e,2} - B_{e,1}]
\end{align*}
\]

\( \Delta F_\text{f} \) quantifies the effect of the change in absorptivity
and \( \Delta F_\text{ze} \) the effect of the change in emission height. If
\( F, B, T \) and \( B_{e,1} \) are known, \( B_{e,2} \) and \( B_{e,1} \) can be computed
using Eq. 21, and therefore Eqs. 22b and 22c can be used to compute \( \Delta F_\text{f} \) and \( \Delta F_\text{ze} \). Therefore, any radiative
code, no matter how complex, that computes \( F, B, T \) and \( B_{e,1} \),
which is generally the case, can be used to compute the changes \( \Delta F_\text{f} \) of the flux at the tropopause that is due to
the change in absorptivity and the change \( \Delta F_\text{ze} \) that is due
to the change in emission height. However, the change in
emission height itself is more difficult to compute as it re-
quires the use of Eq. 14, which is not straightforward for
many radiative codes. This is one of the reasons why we
used a radiative code based on the Net Exchange Formalism
(NEF) (Green, 1967; Cherkaoui et al., 1996; Dufresne
et al., 2005).

\[ b. \ \text{A reference line by line model based on a Net Exchange}
\text{Formulation} \]

The line by line radiative model we use is presented
in Eymet et al. (2016) and its main originality is to
rely on the Net Exchange Formalism (NEF). In a first
step (Kspectrum code), a synthetic high-resolution (typi-
cally 0.0005cm\(^{-1}\)) absorption spectra is computed for
the required atmospheric profile using the HITRAN 2012
molecular spectroscopic database (Rothman et al., 2013)
with Voigt line profiles. For CO\(_2\), sub-lorentzian correc-
tions are taken into account. For H\(_2\)O, the CKD contin-
uum is used with a 25cm\(^{-1}\) truncation and removing the
“base” of each transition (Clough et al., 1989; Mlawer
et al., 2012). In a second step (HR_PPart code), radiative
transfer is computed based on 1D (over a single line of
sight) or 3D (angularly integrated) analytical expres-
sions of spectral radiative Net Exchange Rates and spec-
tral radiative fluxes. We compute the radiative forcing for
CO\(_2\) and H\(_2\)O changes based on the experiments defined
in Collins et al. (2006): the reference experiment, which is
the mid latitude summer (MLS) atmospheric profile with a
CO\(_2\) concentration of 287ppmv (called 1a), an experiment
where the CO\(_2\) concentration is doubled (called 2b), and
an experiment where the CO\(_2\) concentration is doubled and the concentration of H\(_2\)O is increased by 20% (called
4a). In this example, the only absorbing gases consid-
ered are H\(_2\)O, CO\(_2\) and ozone and the troposphere is dis-
cretized into 31 vertical layers. The results compare well
with those published by Collins et al. (2006), as shown on
Table 1.

c. Results for a realistic atmospheric profile

We use the same mid latitude summer atmospheric prof-
ile and consider only the troposphere, from the surface
\( P_s = 1013\text{hPa}, H_s = 0\text{m}, T_s = 294\text{K} \) to the tropo-
pause \( P_t = 190\text{hPa}, H_t \approx 12.6\text{km}, T_t = 218.4\text{K} \), as presented
above.

We first focus on two CO\(_2\) weak absorbing lines. In
Fig. 10, and only in this figure, we exclude absorption by
the H\(_2\)O continuum in order to have an optical thickness
that is as small as possible. For the weaker absorption
line for which the optical thickness is always less than one
(Fig. 10, left column), the shape of the optical thickness
resembles that of the idealized one (Fig. 7). The optical
thickness is low \( (\tau_s \ll 1) \) and the emission height is
about 2-3 km, as expected from Fig. 9. When doubling the
CO\(_2\) concentration, the change in optical thickness is
almost equal to the value for the reference atmosphere, the
difference is due to some absorption by H\(_2\)O. The change
in emission height is less than 100m at wavenumbers far
away from the absorption line center and increases to a few
hundred meters at the absorption line center. The change
in the tropopause irradiance is largely dominated by the
contribution of the change in absorptivity.

For a more absorbing line with a companion weak ab-
sorbing line (Fig. 10, right column), the emission height
is about 2-3 km far from the absorption line center, where
the optical thickness is below one. At the absorption line
center, the emission height reaches 8 km, which is closer
to the tropopause. When doubling the CO\(_2\) concentration,
the change in emission height is a few hundred meters far
from the absorption line center to more than a kilometer at
the absorption line center. The change in the tropopause
irradiance is dominated by the contribution of the change
in emission height.

The results we obtained with the various idealized con-
figurations are consistent with those we obtained with the
reference model. The understanding we gained with the
idealized examples can be applied to interpret the results
with much more complex and realistic models.

We now consider the “thermal infrared” spectral inter-
val from 100 to 2500 cm\(^{-1}\) (4 to 100 \(\mu\text{m}\)). On Fig. 11, 12
and 14, variables are smoothed on a 10cm\(^{-1}\) spectral inter-
val to make the figure more readable. The spectral depen-
dency of the radiative flux at the TOA, at the tropopause,
within the atmosphere, and of the radiative cooling rate in
the atmosphere, as well as how they change when chang-
ing the CO\(_2\) concentration have already been addressed in
many studies (Kiehl and Ramanathan, 1983; Kiehl, 1983;
Charlock, 1984; Clough and Iacono, 1995; Harries et al.,
2001; Huang, 2013). Mlynczak et al. (2016) show that
these results were remarkably insensitive to known uncertainties in the main CO₂ spectroscopic parameters. Zhong and Haigh (2013) showed how the flux at the TOA varies over a wide range of CO₂ values, and they showed that the spectral response is very different depending on the CO₂ concentration. Here we consider the response for an atmosphere with a CO₂ concentration close to its preindustrial value (287ppmv).

The total optical thickness τ, (Fig. 11-a, black line) is primarily due to H₂O absorption, except around 660 and 2300 cm⁻¹ where the two CO₂ strong absorption band systems at 15μm and 4.3μm (magenta) are dominant. The total optical thickness varies over many orders of magnitude, from about one in the atmospheric window (between 800 and 1200 cm⁻¹) to 10⁴ - 10⁵ in the H₂O and CO₂ absorption bands. When the data is not smoothed, the range is even larger, from a few tenths up to 10⁶.

The emission height (Fig. 11-b) almost increases with the logarithm of the total optical thickness τ, (Huang and Bani Shahabadi, 2014). It varies for 2km in the atmospheric window up to 12km, i.e. almost the tropopause height, in spectral region where the optical thickness is very high, especially for the CO₂ bands. For the same optical thickness, the emission height in the CO₂ absorption bands are larger than for the H₂O absorption bands as the CO₂ concentration is uniform over the whole troposphere whereas the H₂O concentration strongly decreases with height.

We define the emission temperature as the temperature for which the Planck function is equal to B(τ), defined by Eq. 20b. The emission temperature (Fig. 11-c) directly follows the evolution of the emission height. The dependence is about 7K/km, as one may expect from the value of the temperature gradient in the troposphere. The upward flux at the tropopause may have been emitted either from the surface or from the troposphere (Eq. 17 and 6). One can see in Fig. 11-d that almost all the flux at the tropopause has been emitted by the troposphere, except in the atmospheric window where both the emission by the surface and the troposphere contribute almost equally (Costa and Shine, 2012).

The ozone absorption band around 1050cm⁻¹ has a specific signature as ozone is mainly located in the higher part of the troposphere. This band has little impact on the optical thickness but has a visible signature on the emission height, the emission temperature, and the outgoing flux (Fig. 11).

Table 1. Difference of the net flux (in Wm⁻²) at the TOA, at 200 hPa (ΔF(200)) and at the surface (ΔFᵣ) for a CO₂ doubling and for an increase of H₂O by 20% for the MLS atmospheric profile and for the thermal infrared spectral interval (100-2500 cm⁻¹, i.e. 4 to 100 μm). The results computed with our model (KS) are compared to those published by Collins et al. (2006) (C06) for an ensemble of line by line models using the same atmospheric profiles (see Sect. 4-b).

<table>
<thead>
<tr>
<th>experiments</th>
<th>ΔF_TOA</th>
<th>ΔF(200)</th>
<th>ΔFᵣ</th>
</tr>
</thead>
<tbody>
<tr>
<td>2×CO₂ (2b-1a)</td>
<td>2.81</td>
<td>5.57</td>
<td>1.67</td>
</tr>
<tr>
<td>1.2×H₂O (4a-2b)</td>
<td>3.71</td>
<td>4.60</td>
<td>1.14</td>
</tr>
</tbody>
</table>

Fig. 12 displays the changes in total optical thickness, emission height, emission temperature and upward flux at the tropopause when doubling the CO₂ concentration or when increasing the H₂O concentration by 20%. For the CO₂ 15μm band system (660cm⁻¹), the change in emission height, emission temperature, and tropopause flux is maximum on the edges of the band (Fig. 12-bcd), where the CO₂ optical thickness is about a few units (Fig. 11-a). In these spectral regions, the change in emission height is about 1km and the change in emission temperature is about 7K. The change in emission height is almost zero at the band center as the emission height is already close to the tropopause, i.e. close to the maximum height. The change in the flux at the tropopause is almost only due to the change in emission height (Fig. 12-d). For the 4.3μm (2300 cm⁻¹) CO₂ band, the changes in emission height and emission temperature resemble those for the 15μm (660cm⁻¹) band, but these changes have almost no impact on the tropopause flux as the Planck function is almost zero at these wave numbers for the atmospheric temperature. In addition to these two very strong absorption bands, CO₂ also has some minor bands that produce small changes in emission height, emission temperature and tropopause flux. In the spectral domain of these minor bands, the optical thickness is small (about 10⁻³) and is due to absorption by both H₂O and CO₂. As a result, both the change in emission height and in absorptivity play a comparable role, whereas the change in absorptivity would have had a dominant role if CO₂ were the only absorbing gas. Note that this holds for the current atmosphere but not for an atmosphere with very high CO₂ concentration: these “minor” bands contribute to the CO₂ forcing by about 6% in current conditions, but they contribute by about 25% for CO₂ concentration that are 100 times larger (Augustsson and Ramanathan, 1977; Zhong and Haigh, 2013).

As the change ΔZₑ in emission height strongly varies with wavenumber, we define its average value in two ways. The first is the broadband average <ΔZₑ>, where ΔZₑ is weighted by the Planck function at surface temperature, as for the broadband absorptivity shown on Fig. 1-a. We found a value of 150m, exactly as Held and Soden (2000). As explained in this article, the broadband change
in emission height can be directly used to compute the radiative forcing. However, the change in the flux at the tropopause is different from zero only in limited spectral regions where \( \Delta Z_e \) is also large (Fig. 12). Therefore we define a second average, \( < \Delta Z_e >_F \), namely the “forcing average” change in emission height defined as the average of \( \Delta Z_e \) weighted by \( \Delta F_{Z_e} \):

\[
< \Delta Z_e >_F = \frac{\int_{0}^{\infty} \Delta Z_e(\nu) \Delta F_{Z_e}(\nu) d\nu}{\int_{0}^{\infty} \Delta F_{Z_e}(\nu) d\nu} \quad (23)
\]

This quantity is the change in emission height that actually contributes to the radiative forcing. We obtain a value of 1025 m, which is much larger than the broadband mean. The change in CO\(_2\) concentration impacts the flux at the tropopause in the very few spectral regions where the optical thickness of the atmosphere is about a few units. In these spectral regions the change in emission height is on average 1025 m. The mean emission height itself is less sensitive to the average method: The broadband emission
height is 5800m whereas the “forcing average” emission height is 6100m.

When the H₂O concentration is increased, the change in emission height is about 200 m (Fig. 12-b) over spectral intervals that are much wider (100-600 cm⁻¹, 1300-2000 cm⁻¹) than for CO₂. In these intervals the absorption by H₂O is strong and the change of the flux at the tropopause is almost only due to the change in emission height (Fig. 12-e). In spectral regions where absorption by CO₂ dominates (600-750 cm⁻¹), the change in H₂O is completely masked by the CO₂ absorption. In most of the atmospheric window (750-1300 cm⁻¹), the change in emission height is small (< 100m) and the change of the flux at the tropopause is mainly due to the change in absorptivity, with a significant contribution of the water vapor continuum (Costa and Shine, 2012). An exception is around 1050 cm⁻¹ where ozone absorbs. In this spectral region both the ozone and the water vapor emit radiation and the emission height includes both the contribution of ozone, which is mainly located in the high troposphere, and the contribution of water vapor, which is mainly located in the lower troposphere. When the H₂O concentration increases, the radiation emitted by H₂O that reaches the tropopause increases whereas the radiation emitted by ozone that reaches the tropopause does not change. As
a result the emission height decreases by about 200m (Fig. 12-b), the emission temperature increases (Fig. 12-c) and the contribution of the change in emission height to the flux at the tropopause is positive (Fig. 12-e).

When considering the radiative flux over the whole thermal infrared domain, the decrease of the flux at the tropopause due to an increase of CO₂ is primarily due (by about 90%, Table 2) to the change in emission height, the change in absorptivity playing a minor role (about 10%). For an increase of water vapor, the change in absorptivity plays a more important role (about 40%) but the change in emission height still plays the dominant role (≈ 60%). However, this significant contribution of the change in absorptivity for H₂O is primarily due to the H₂O continuum. When the continuum is suppressed, the change in emission height is as high as 80% and the contribution of the change in absorptivity reduces to 20%.

**Table 2.** Difference Δ\(\bar{F}\) of the upward flux at the tropopause, difference Δ\(\bar{F}_{Ze}\) of this flux due to change in emission height, difference Δ\(\bar{F}_{\gamma}\) due to change in absorptivity, relative contribution of each of the changes (Δ\(\bar{F}_{Ze}\) and Δ\(\bar{F}_{\gamma}\)) to the total Δ\(\bar{F}\), and change Δ\(\bar{F}_s\) of the broadband transmissivity of the atmosphere. The differences are computed for a CO₂ doubling (2b-1a, first row) and an increase of H₂O by 20% (4a-2b, second row), for the atmospheric profiles presented in the text (Sect. 4-b).

<table>
<thead>
<tr>
<th>atm. profile</th>
<th>experiments</th>
<th>Δ(\bar{F}) (Wm⁻²)</th>
<th>Δ(\bar{F}_{Ze}) (Wm⁻²)</th>
<th>Δ(\bar{F}_{\gamma}) (Wm⁻²)</th>
<th>Δ(\bar{F}_{Ze}/\bar{F}) (-)</th>
<th>Δ(\bar{F}_{\gamma}/\bar{F}) (-)</th>
<th>Δ(\bar{F}_s) (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MLS</td>
<td>2×CO₂ (2b-1a)</td>
<td>-3.83</td>
<td>-2.19</td>
<td>-2.19</td>
<td>0.00</td>
<td>0.00</td>
<td>-6.30 × 10⁻³</td>
</tr>
<tr>
<td></td>
<td>1.2×H₂O (4a-2b)</td>
<td>-3.78</td>
<td>-2.21</td>
<td>-2.21</td>
<td>0.00</td>
<td>0.00</td>
<td>-2.70 × 10⁻²</td>
</tr>
</tbody>
</table>

**Fig. 13.** Temperature as a function of altitude for the full MLS atmospheric profile, i.e. including the stratosphere. Both the reference temperature (dash line) and the temperature after the stratosphere has adjusted to a doubling of the CO₂ concentration (continuous line) are shown.

**Fig. 14.** For the MLS atmospheric profile including the stratosphere, emission height \(Z_e\) (a), radiative flux at the TOA (b) total (black), emitted by the surface (blue) and emitted by the atmosphere (red), change in emission height (c) (Δ\(Z_e\), magenta) and in the flux at the TOA (d) if the temperature in the stratosphere is held fixed (dash line) or is adjusted (continue line). The abscissa is given in wave-number (cm⁻¹) at the bottom and in wavelength (\(\mu m\)) at the top. Variables are smoothed on a 10cm⁻¹ spectral interval.

**d. Including the stratosphere**

So far and for simplicity we considered an atmosphere that extends from the surface to the tropopause, and therefore in which the vertical temperature gradient is always negative and driven by the convective adjustment. We now consider an atmosphere that extends to an altitude...
of 100 km and will show that the main results are still valid when the temperature adjustment of the stratosphere is taken into account.

In the stratosphere, the radiative cooling is compensated by the dynamic heating with a relaxation time of a few months. As the dynamics in the troposphere and in the stratosphere are weakly coupled, it has been shown (Hansen et al., 1981, 1997; Stuber et al., 2001; Forster et al., 2007) that it is more relevant to compute the radiative forcing after allowing stratospheric temperatures to adjust to a new radiative equilibrium than to compute the radiative forcing with a fixed stratospheric temperature. The stratospheric temperature adjustment is computed assuming no change in stratospheric dynamics as follows: after computing the radiative budget $S_1(z)$ at each altitude $z$ for the reference concentration and temperature, the radiative budget $S_2(z)$ at each altitude $z$ is computed with the same temperature profile but a modified CO$_2$ concentration. The temperature in the stratosphere is then adjusted until $S_2(z) \approx S_1(z)$ at each altitude $z$ of the stratosphere. The results we obtain for the MLS profile and a doubling of the CO$_2$ concentration are shown in Fig. 13. By construction the temperature in the troposphere does not change. The temperature in the stratosphere decreases as expected (Hansen et al., 1997; Stuber et al., 2001), with a temperature cooling of 5 to 10K.

Compared to the troposphere only case (Fig. 11), including the stratosphere increases the emission height in the center of the CO$_2$ absorption band systems at 15 and 4.3 µm where the emission height reaches values up to 30 to 40 km (Fig. 14-a). This height is even larger when looking at the full resolution data (not shown). Including the stratosphere increases the optical thickness around the 9.7 µm ($\approx 1050$ cm$^{-1}$) O$_3$ absorption band by a few units, which has a significant impact on both the emission height and the flux at the TOA (Fig. 14-a and b).

When doubling the CO$_2$ concentration, the change in emission height (Fig. 14-c) is comparable to the case without stratosphere (Fig. 12-b) except in the 15 and 4.3 µm CO$_2$ absorption bands. At these band centers, the emission height can now be larger than the tropopause height, the increase in emission height is not blocked anymore and it has an almost constant value of about 3 km. One can show that the change in emission height is almost constant for a well-mixed absorption gas when absorption is saturated because the emission height is then close to the height where the optical thickness is equal to one (Fig. 6). This large change in emission height has a clear signature on the change of the flux at the TOA for the 15 µm CO$_2$ absorption band. At the absorption band center, a higher emission height leads to an increase of the outgoing flux because the temperature vertical gradient in the stratosphere is positive. However, this happens only if the temperature in the stratosphere is fixed (Fig. 14-d, dash line) as already shown (Kiehl, 1983; Charlock, 1984). If the temperature of the stratosphere is adjusted as explained above, the decrease of temperature in the stratosphere leads to a decrease of the emitted radiation. As a result, the change in the outgoing flux at the center of the 15 µm CO$_2$ absorption band is slightly negative. The pattern of the change of the spectral flux around the 15 µm CO$_2$ absorption band is similar if the atmosphere only extends up to the tropopause (Fig. 12-d) and if the atmosphere extends higher than the tropopause but the stratospheric adjustment is considered (Fig. 14-d, continuous line). At the first order, the interpretation of the results we obtained with an atmosphere reduced to the troposphere can be extended to a full atmosphere where the temperature of the stratosphere is adjusted. However, the adjustment of the stratosphere also impacts the emission by other gases: H$_2$O for wavenumbers lower than 500 cm$^{-1}$ and ozone near 1050 cm$^{-1}$.

5. Summary and conclusion

In this article we presented a framework that allows us to make a direct and precise link between the basic radiative transfer equations in the atmosphere on one hand, and the concept of emission height on the other hand. This allowed us to quantify how much a change in the greenhouse effect originates from a change in the emission height and how much originates from a change in the absorptivity of the atmosphere, i.e. the absorption over the entire height of the atmosphere.

The fact that a saturation of the absorptivity of the atmosphere leads to a saturation of the greenhouse effect is directly related to the hypothesis of an isothermal atmosphere. When this simplification is removed and the decrease of temperature with altitude is considered, as it is the case in the troposphere, the greenhouse effect can continue to increase even if the absorptivity of the atmosphere is saturated.

The fundamental difference between our approach and other approaches such as the “bulk emission temperature” (Benestad, 2017) or the “brightness temperature” commonly used in remote sensing, is that we split the radiation leaving the atmosphere toward space in two terms: the radiation that has been emitted by the surface (termed surface transmitted irradiance in Costa and Shine (2012)), and the radiation that has been emitted by the atmosphere. The fraction between these two terms is directly driven by the absorptivity of the atmosphere (Eq. 8a). When the absorptivity is zero, the total flux leaving the atmosphere originates from radiation emitted by the surface, and the atmosphere has no radiative impact. When the absorptivity is close to 1, i.e. when the total optical thickness of the atmosphere is larger than about 4, the opposite situation happens: the total flux leaving the atmosphere has been emitted by the atmosphere, the surface does not have
any direct radiative impact on the flux leaving the atmosphere, and increasing the optical thickness does not have any influence on the ratio between these two terms anymore. However, this does not mean that the greenhouse effect does not change. Increasing the optical thickness increases the mean emission height and if the atmosphere is not isothermal, a change in emission height translates in a change in outgoing radiative flux.

For an increase in CO$_2$ concentration above its preindustrial value, the increase of the greenhouse effect is primarily due (by about 90%) to the change in emission height. In spectral regions that actually contribute to the radiative forcing, the increase in emission height is about 1 km for a doubling of the CO$_2$ concentration. As the mean emission height is about 6 km, i.e. above where most of the mass of water vapor is located, the radiative effect of this change of emission height is weakly affected by the water vapor amount. This explain why the increase of the greenhouse effect when increasing CO$_2$ is weakly dependent of the H$_2$O amount (Fig. 1-b), in contrast with the broadband absorptivity. The change in emission height will be of comparable magnitude for any other well mixed absorbing gases in the spectral domains where the absorptivity is saturated. For an increase of water vapor, the change in absorptivity plays a more important role (about 40%) but the change in emission height is still about 60%. Indeed, away from the atmospheric window, the absorptivity by water vapor becomes saturated and the change in emission height becomes therefore dominant.

The emission height depends on both the temperature profile and the optical properties (Eqs. 14 and 11). We showed that the classical assumption that the emission height is close to the altitude where the optical thickness between this altitude and the top of the atmosphere is equal to one is valid only for atmospheres that are optically thick enough ($\tau_s \gg 1$). For optically thin atmospheres or with and optical thickness close to one, this assumption is not valid and leads to an underestimation of the emission height.

Considering the real temperature vertical profile in the whole atmosphere makes simplified analysis of the greenhouse effect a priori difficult. However, this complexity is essentially eliminated when considering the adjustment of the stratospheric temperature. This has long been shown when considering global fluxes. Here, we have shown that this is also the case when looking at the change in spectral fluxes and emission altitude, and therefore that it is legitimate to replace the vertical profile of the entire atmosphere by the vertical profile of the troposphere alone, for simplified thinking.

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APPENDIX

A1. Analytical expression for the idealized atmosphere

In this section we take advantage of the assumption of the idealized atmosphere (section 2) and in particular that the spectral Planck function increases linearly with pressure (Eq. 4).

a. Outgoing radiance $\mathcal{F}$ at the tropopause

For the idealized atmosphere, Eq. 8b can be integrated analytically and one obtains (after an integration by parts):

$$\mathcal{B}_e = \mathcal{B}(P_s) + \left[\mathcal{B}(P_t) - \mathcal{B}(P_s)\right]\left[\frac{1}{1 - e^{-\tau_t}} - \frac{1}{\tau_s}\right]$$

(A1)

The outgoing radiance $\mathcal{F}$ at the tropopause is (Eq. 6):

$$\mathcal{F} = \mathcal{B}_s e^{-\tau_s} + (1 - e^{-\tau_t}) \mathcal{B}_e$$

(A2)

If the atmosphere is optically very thin ($\tau_s \ll 1$), one may obtain that $\mathcal{B}_e \approx \left[\mathcal{B}(P_t) + \mathcal{B}(P_s)\right]/2$ and therefore $\mathcal{F} \approx \mathcal{B}_s + \tau_t [\mathcal{B}_e - \mathcal{B}_s]$. The outgoing radiance is the same for the isothermal and non-isothermal atmosphere, it is equal to the radiance $\mathcal{B}_s$ emitted by the surface when the atmosphere is perfectly transparent ($\tau_s = 0$), and then decreases linearly with $\tau_s$ when the latter increases. In contrast, if the atmosphere is optically very thick ($\tau_s \gg 1$), $\mathcal{F} \approx \mathcal{B}_e \approx \mathcal{B}(P_t)$, the outgoing radiance is equal to the radiance emitted by a black-body, which temperature is that at the tropopause.

b. Weight $\omega(P)$

$\omega(P)$ is the normalized weighting function to compute the equivalent blackbody emission of the atmosphere (Eq. 8b). Note that according to Eq. 8c, $\int_{P_t}^{P_s} \omega(P)dP = 1$, and $-\omega(P)$ can be interpreted as a probability density function. For the idealized atmosphere, an according to Eqs. 5, 7 and 8c

$$\omega(P) = \frac{1}{1 - \tau_s} \left(\frac{-k_f}{g}\right) e^{-k_f(P-P_t)/g}$$

(A3)

1 a second order Taylor development is required for $1/(1 - e^{-\tau_t})$
If the atmosphere is optically thin, \( \tau_s = k f (P_s - P_t)/g \ll 1 \), and after a first order Taylor expansion one obtains:

\[
\omega(P) \approx -\frac{1}{P_s - P_t} \quad (A4)
\]

If the atmosphere is optically thin, the weight \( \omega \) is constant along the vertical.

If the atmosphere is optically thick, \( \tau_s = \kappa (P_s - P_t) \gg 1 \), \( \tau_s \approx 0 \) and, according to Eq. A3, \( \omega \approx 0 \) in the lower troposphere, and \( \omega \approx -\frac{k f}{\kappa} \) close to the top of the atmosphere.

### A2. Difference in TOA flux when changing the atmospheric absorption

The difference of the flux at the TOA for two atmospheres that only differ by their absorbing gases can be written using Eqs 21 and 20b as:

\[
F_2 - F_1 = \left( \tilde{\tau}_{s,2} - \tilde{\tau}_{s,1} \right) B_s \quad (A5)
\]

\[
+ (1 - \tilde{\tau}_{s,2}) \int_{P_t}^{P_s} \omega_2(P) B(P) dP
\]

\[
- (1 - \tilde{\tau}_{s,1}) \int_{P_t}^{P_s} \omega_1(P) B(P) dP
\]

which can be written as:

\[
F_2 - F_1 = \left( \tilde{\tau}_{s,2} - \tilde{\tau}_{s,1} \right) B_s \quad (A6)
\]

\[
+ (1 - \tilde{\tau}_{s,2}) \int_{P_t}^{P_s} (\omega_2(P) - \omega_1(P)) B(P) dP
\]

\[
- (\tilde{\tau}_{s,2} - \tilde{\tau}_{s,1}) \int_{P_t}^{P_s} \omega_1(P) B(P) dP
\]

and finally as:

\[
F_2 - F_1 = \left( \tilde{\tau}_{s,2} - \tilde{\tau}_{s,1} \right) [B_s - B_{e,1}] \quad (A7)
\]

### References


