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A Colorful look at Climate Sensitivity

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Abstract. The radiative response to warming, and to changing concentrations of CO_2, is studied in spectral space. If relative humidity does not change with temperature, clear-sky emissions over spectral intervals in which water vapor is optically thick become independent of surface temperature, giving rise to the idea of spectral masking. It is demonstrated that this idea allows one to derive simple, physically informative, and surprisingly accurate, expressions for the clear sky radiative forcing, radiative response to warming and hence climate sensitivity. Extending these concepts to include the effects of clouds, leads to the expectation that (i) clouds damp the clear-sky response to forcing, (ii) that diminutive clouds near the surface, which are often thought to be unimportant, may be particularly effective at enhancing the clear-sky sensitivity over deep moist tropical boundary layers; and (iii) even small changes in high-clouds over deep moist regions in the tropics makes these regions radiatively more responsive to warming that previously believed. The analysis demonstrates that the net effect of clouds on warming is ambiguous, justifying the assertion that the clear-sky (fixed RH) climate sensitivity – which after accounting for clear-sky surface albedo feedbacks, is about 3 K – provides a reasonable prior for Bayesian updates accounting for how clouds are distributed, how they they might change, and for deviations associated with changes in relative humidity with temperature. These effects are best assessed by quantifying the distribution of clouds and water vapor, and how they change, in temperature, rather than geographic space.

1 Introduction

In recent years a spectral treatment of thermal infrared radiation has helped advance understanding of many basic aspects of Earth's energy balance and how it responds to forcing. For instance, a consideration of the differential spectral response of outgoing long-wave radiation to warming has proved crucial to understanding the linearity of outgoing OLR (Koll and Cronin, 2018). A spectral treatment of thermal-infrared radiation also helps explain why the Planckian response to warming is susbstantially smaller than would be expected from the Stefan Boltzmann law (Cronin and Dutta, 2022). Likewise the near uniformity of clear-sky radiative cooling (Jeevanjee and Fueglistaler, 2020), and Earth's ultimate ability to radiatively respond to forcing at very high temperatures proves difficult to explain without a consideration of the spectral signature of absorbers (Kluft et al., 2021; Seeley and Jeevanjee, 2021). Many of the above studies helped answer important questions by abandoning the idea that atmospheric radiative transfer could usefully be thought about as broadband, or grey. Grey atmospheres don't allow for spectral masking, which is a central concept upon which much of the emergent understanding has been built.

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In what follows, this more colorful way of thinking about radiative transfer is developed further, to show that a spectral view of radiative transfer, and in particular spectral masking (a concept defined more precisely below), helps quantify Earth's equilibrium climate sensitivity. Armed with spectral thinking, the clear-sky climate sensitivity can – to a reasonable degree of approximation – be derived analytically. This provides a firmer foundation for understanding how clouds and other trace-gases modify Earth's response to *forcing*[1]. A foundation on which it also becomes easier to understand, and eventually quantify how clouds, in particular, even with out changing, mediate Earth's response to *warming*. In addition to explaining behaviors which were previously known, but not well understood, new and sometimes surprising properties of Earth's greenhouse atmosphere are uncovered.

The ideas presented here were developed in lectures on the greenhouse effect the first author gave at the Universität Hamburg, in the Fall of 2021. Many had their origins in joint work with the second author. Subsequently we became aware that we were not the only ones thinking along these lines. The simple model of CO_2 forcing presented in those lectures, was developed independently and much earlier by Wilson and Gea-Banacloche (2012), and was elaborated upon further and more thoroughly by Seeley (2018) and Jeevanjee et al. (2021b). Likewise as we later learned, the ideas related to the clear-sky radiative response were being developed simultaneously by others. Different elements of this more colorful way of thinking have recently appeared in papers by Jeevanjee et al. (2021a); McKim et al. (2021); Colman and Soden (2021) and continue to be developed by these authors. At the risk of repetition, we present our ideas as we originally developed them, as they are foundational for novel arguments as to the role of clouds in Earth's radiative response to forcing, and the implications this has for assessing climate sensitivity more broadly.

The outline of the paper is as follows, after introducing the data sources and community tools used, the basic ideas behind spectral masking are introduced in §3, and used to derive estimates and provide understanding of Earth's clear-sky climate sensitivity and its components in §4. This provides a basis for thinking about Earth's equilibrium climate sensitivity more broadly (§5), and for better understanding the role of clouds in its determination. Conclusions and an outlook are presented in §6

2 Data

Absorption spectra of selective absorbers, here CO_2 and H_2O are taken from the catalog used for the Atmospheric Radiative Transfer Simulator, ARTS (Buehler et al., 2018; Eriksson et al., 2011). ARTS includes treatments of line broadening – with the treatment of the foreign-broadening appropriate for Earth's atmosphere, and a representation of continuum absorption following the approach of Clough et al. (1989, 2005) as modified by Mlawer et al. (2012). Other data sources include monthly mean, gridded ($0.25° \times 0.25°$) near surface ($2\,\mathrm{m}$) air temperatures and column water vapor for the 240 months between 2001 and 2021, and are taken from reanalyses of meteorological data (ERA5) (Hersbach et al., 2019). Cloud data is based on

[1]Here forcing is used generically, for instance to refer to a change of atmospheric composition, and distinguished from *radiative forcing*, which is the response.

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Preprint. Discussion started: 2 January 2023

measurements using the AATSR instrument which flew on ENVISAT (Poulsen et al., 2019). The record extends from 2002-05 through 2012-04 and level 3 cloud-top temperature and cloud fraction are used.

3 Theory

Concepts are developed for understanding the emission of terrestrial radiation, $99\,\%$ of which is emitted in the $50\,\mathrm{cm}^{-1}$ to $2000\,\mathrm{cm}^{-1}$ wave-number interval, which is sometimes referred to as the long-wave or thermal infrared part of the electromagnetic spectrum. Earth's atmosphere is assumed to consist of a background gas (dry-air) that is fully transparent to radiation emitted from its surface, absorption is limited to minor constituents, either particulate (clouds) which are assumed to be broadband (grey) absorbers, or by trace amounts of greenhouse gases which act as selective (or colorful) absorbers. A differentiation is made between short and long-lived greenhouse gases, whereby the phrase 'short-lived' is short-hand for gases, namely water vapor, which are assumed to exist in abundances (partial pressure) determined by the atmospheric temperature. Long-lived gases, like CO_2, are well mixed, and assumed to be independent of the ambient temperature. Scattering by terrestrial radiation is not considered, to the extent solar radiation is considered clouds are treated as conservative scatterers, and gaseous absorption is neglected.

Our arguments are based on three simplifications:

- **S1** The effects of continuum absorption and line-broadening, while resolved spectrally, can be approximated by their values at an effective pressure, P (and in some cases temperature T, and composition) taken to be representative of the entire column.
- **S2** The atmospheric relative humidity, $\mathcal{R}$ is a fixed function of temperature, i.e., $\mathcal{R}(T)$.
- **S3** At a given wave-number, ν, *changes* in emission to space from a blackbody are assumed to be attenuated by the optical path above it.

S1 is mostly made for convenience, as it makes the physical arguments more transparent. In a somewhat stronger form (constant $\mathcal{R}$) the implication of **S2** was exploited by Nakajima et al. (1992); Ingram (2010) whose ideas have been foundational, as well as by Goldblatt et al. (2013); Jeevanjee et al. (2021a), and in the earlier studies of runaway greenhouse atmospheres by Komabayasi (1967); Ingersoll (1969). **S3** differs from Beer's law, which would normally be dismissed out of hand as it neglects the emission of the absorber, in that it refers to the *changes* in emission. **S1**-**S3** are justified in more detail below, and their implications for understanding and quantifying Earth's equilibrium climate sensitivity are explored.

3.1 Fixed relative humidity

3.1.1 Theory

The statement that the relative humidity, $\mathcal{R}$, does not change with warming (Arrhenius, 1896; Manabe and Wetherald, 1967) contains a subtle ambiguity. Is $\mathcal{R}$ constant as a function of height, z, atmospheric pressure, P, or temperature T? For a com-

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

pressible atmosphere all three cannot be true. On physical grounds there is a reason to expect that $\mathcal{R}$ is more constant with height in the lower troposphere (for instance the relative humidity over the sea-surface varies little over large ranges of temperature) and more constant with temperature in the upper troposphere, which deepens with warming. **S2** adopts the latter description which implies that $P_{\mathrm{v}}(T) = \mathcal{R}(T)P_*(T)$, where P_{v} and P_* denote the partial pressure of water vapor and its saturation value respectively.

Subject to weak restrictions on how T changes with P, a consequence of the above is that the water vapor burden,

$$W(z) = \int_z^\infty \rho_{\mathrm{v}}\,\mathrm{d}z \tag{1}$$

at some height z, with $\rho_{\mathrm{v}}(z)$ the vapor density, is determined by the temperature, T, at that height alone. To show this we assume a hydrostatic atmosphere, $\rho g\,\mathrm{d}z = -\mathrm{d}P$, with ρ the total density of the air, and g the gravitational acceleration, and allow for a coordinate transform from $P \to T$ (which is equivalent to $P(T)$ being single valued) It follows that

$$W(T) = \int_T^{T(\infty)} P_{\mathrm{v}}(T') \left(\frac{R}{gR_{\mathrm{v}}} \frac{\mathrm{d}\ln(P)}{\mathrm{d}T} \right) \mathrm{d}T', \tag{2}$$

with R the mass specific gas constant of the air, and R_{v} the value for water vapor. For the case whereby T follows an unsaturated adiabat corresponding to moist air, $\frac{\mathrm{d}\ln(P)}{\mathrm{d}T} = c_p/R$. In this case and subject to the specific heat of air, c_p, R_{v} and the gravitational acceleration, g, being constant,

$$W(T) = \frac{c_p}{gR_{\mathrm{v}}} \int_T^{T(\infty)} P_{\mathrm{v}}(T')\mathrm{d}T', \tag{3}$$

thereby articulating the conditions under which W is only a function of T.

Among the assumptions used to arrive at Eq. (3) the idea that T and P are in a one-to-one relationship is in most obvious violation of the facts, as while T can (notwithstanding near surface inversions) usefully be approximated as strictly decreasing in the troposphere, it generally increases in the stratosphere. Neglecting this amounts to approximating $W(T)$ as

$$W(T;T_{\mathrm{cp}}) \approx \frac{c_p}{gR_{\mathrm{v}}} \int_T^{T_{\mathrm{cp}}} P_{\mathrm{v}}(T')\mathrm{d}T'. \tag{4}$$

with T_{cp} the temperature of the cold-point tropopause. The effect of this approximation is small, both by virtue of the smallness of $P_{\mathrm{v}}(T_{\mathrm{cp}})$ relative to its values at larger temperatures, and because we are mostly interested in $\mathrm{d}W(T)/\mathrm{d}T$, which is constrained by the smallness of differences in the mass of the stratosphere as the surface warms. Simulations suggests that T_{cp} is effectively constant across a wide range of conditions characteristic of the tropical atmosphere (Seeley et al., 2019). Hence we introduce it as a parameter, with the value $T_{\mathrm{cp}} = 194\,\mathrm{K}$ take from radio occultation measurements in the tropics (Tegtmeier et al., 2020), bearing in mind that the same observations show substantially (20 K) larger values in the extra-tropics.

The idea that $\mathrm{d}\ln(P)/dT$ depends only on T is justified following an unsaturated adiabat, also for moist air. While this is a good approximation for the upper troposphere, it is not justified for the middle and lower troposphere, where the temperature

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

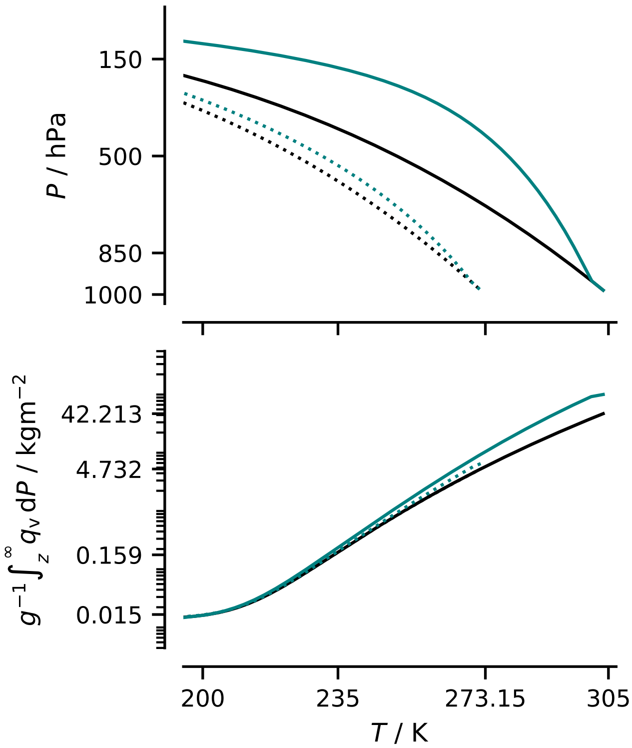

Figure 1. Theoretical temperature profiles and column humidities. Temperature profiles (left) following the formulation of the unsaturated and saturated moist adiabats in Marquet and Stevens (2022) for two different surface pressures and surface humidities. Column water vapor, W, between the top of the atmosphere and the height corresponding to the indicated temperature (right).

more closely follows the isentropic expansion of saturated air. The impact of allowing $\mathrm{d}\ln(P)/dT$ to vary with P as in the saturated case is, following the arguments of Romps (2014), expected however to be small. This is demonstrated in Fig. 1 which compares values of $W(T)$ for temperature following saturated and unsaturated isentropic expansion. These have been calculated for $\mathcal{R} = \text{const.}$. Using a C-shaped profile of $\mathcal{R}$, as is more characteristic of the troposphere (Romps, 2014; Bourdin et al., 2021), albeit modified so the anchoring points depend on T, leads to similar conclusions. In the case when both the C-Shaped profile of $\mathcal{R}$, and the change of the profile of T with P are fit to observations in ways that introduce dependencies other than just on T, W remains well enough approximated by the curves in the right panel of Fig. 2 to support the conceptual arguments.

3.1.2 W–Observations

Support for the theory is provided empirically, based on the co-variability of T_{sfc} and W_{sfc} as estimated by the ERA5 data taken over the period between 2001 and 2021. Fig. 2 summarizes monthly averages. The data demonstrates a relatively strong

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Preprint. Discussion started: 2 January 2023

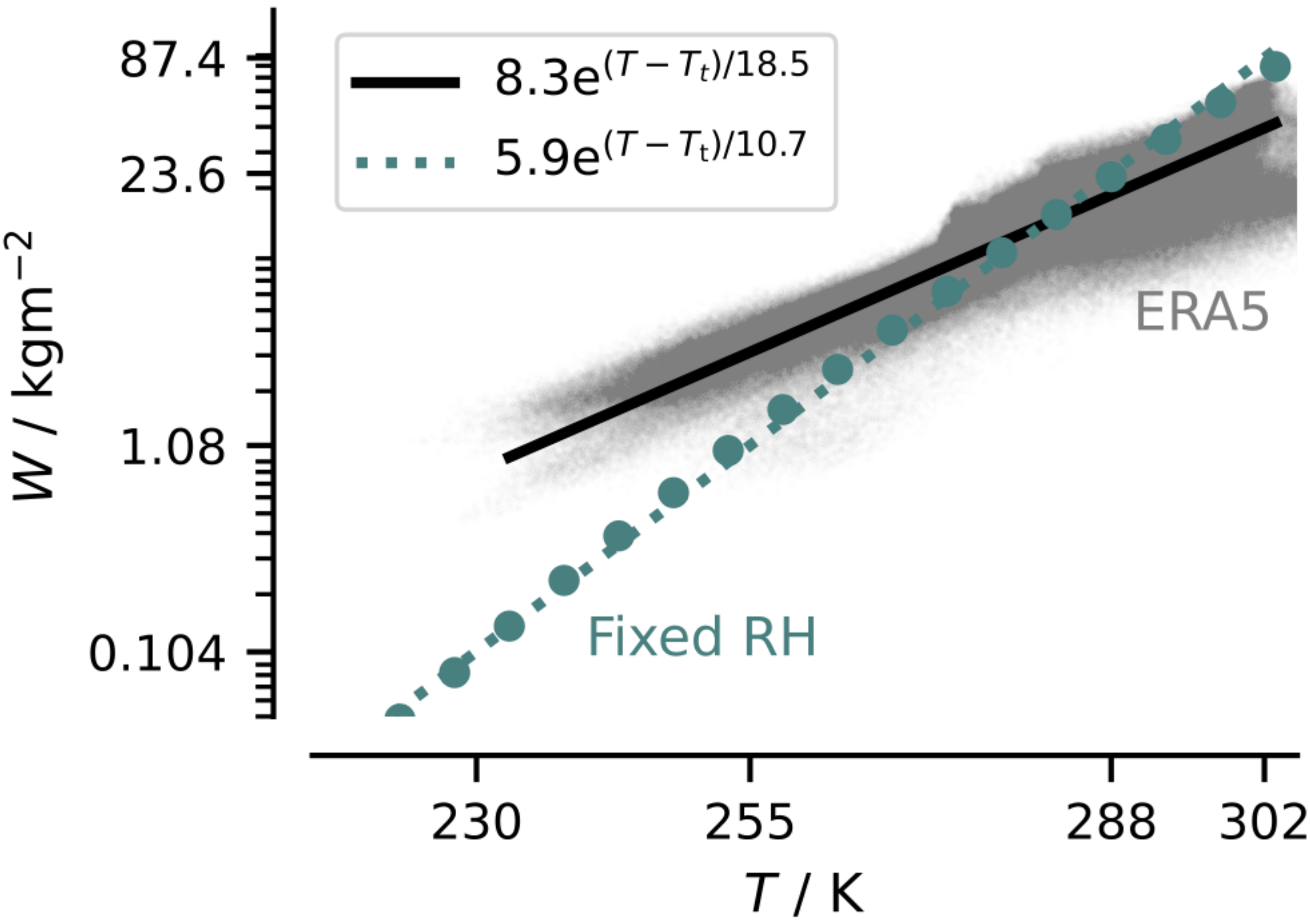

Figure 2. Column water vapor, W, versus T for $T = T_{\mathrm{sfc}}$ and W given by the reanalysis (grey points) and for a fixed $\mathcal{R}(T)$ following an idealized C-shaped profile (filled teal-colored circles). The solid and dotted lines are fits whose slopes are chosen to match those of the grey and teal points respectively, with a crossing point at present-day global temperatures.

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

relationship between $T_{\rm sfc}$ and $W_{\rm sfc}$, which we parameterize as

$$W_{\rm sfc}(T_{\rm sfc}) \approx 8.3 \exp\left(\frac{T - 273.16}{18.5}\right). \tag{5}$$

which is fit to the data by linearly regressing $\ln(W)$ binned by T against T. $\ln W_{\rm sfc} \propto T/18.5$, is flatter than would be expected if every point shared the same $\mathcal{R}$. profile, as is shown by the filled (teal) circles, calculated assuming $\mathcal{R} = 0.8$ (Fig. 2). The best fit to this data, is denoted by $W_{\mathcal{R}}$

$$W(T) \approx W_{\mathcal{R}}(T) = 5.8 \exp\left(\frac{T - 273.16}{10.2}\right). \tag{6}$$

This demonstrates that $\ln W_{\mathcal{R}}$ varies with T nearly twice as strongly as $\ln W_{\rm sfc}$ This difference is not particularly sensitive to how $\mathcal{R}$ is specified, so long as it remains constant – C-shaped profiles yield a similar slope. These findings motivate the rather simple choice of $\mathcal{R} = 0.8$ chosen so that the global mean of W matches the observed global mean. A relative humidity of 0.8 is larger than the mean $\mathcal{R}$, as it must be to capture the non-linearity of $W(T)$, whereby $\overline{W(T)} > W(\overline{T})$, with an over-bar denoting the global average.

The difference between $W_{\mathcal{R}}$ and $W_{\rm sfc}$ is a consequence of the extra-tropics being stormier, and hence relatively more humid, than the tropics, wherein upward moisture transport is intermittent, and broad areas of subsidence desiccate the atmosphere. The atmospheric circulation thus imprints itself on the profile of $\mathcal{R}$: variations of $W_{\rm sfc}$ with $T_{\rm sfc}$ effectively parameterize the circulation regime. To the extent that the atmospheric circulation does not change with warming, one would thus expect the cloud of points, from the distribution of atmospheric profiles, to shift following $W_{\mathcal{R}}$.

3.2 A Simpsonian atmosphere

Following Nakajima et al. (1992) and Ingram (2010) we here review the implications of W depending only on T for the atmospheres optical properties, something Jeevanjee et al. (2021a) calls 'Simpson's Law'.

Assuming water vapor is the only important absorber, the optical depth of the atmosphere between a given height z and space is,

$$\tau_\nu(z, \infty) = \int_z^\infty \kappa_{\nu,\rm v}\, \rho_{\rm v}\, {\rm d}z, \tag{7}$$

with $\kappa_{\nu,\rm v}$ denoting the mass absorption cross section of water vapor at wave-number ν. Assuming $\kappa_{\nu,\rm v}$ doesn't depend on P then the above arguments imply that $\tau_{\nu,\rm v}$ is only a function of T. For the conceptual arguments developed below we make a more extreme assumption, which is that $\kappa_{\nu,\rm v}$ is constant, in which case

$$\tau_\nu(z, \infty) = \tau(T) = \kappa_\nu W(T). \tag{8}$$

In reality $\kappa_{\nu,\rm v}$ does depend on T and P, as both influence line-broadening and empirical corrections for continuum absorption. The justification for using effective absorption coefficients, is that while including the covariability between $\kappa_{\nu,\rm v}$ and $\rho_{\rm v}$ can introduce small quantitative corrections, it adds no conceptual content.

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

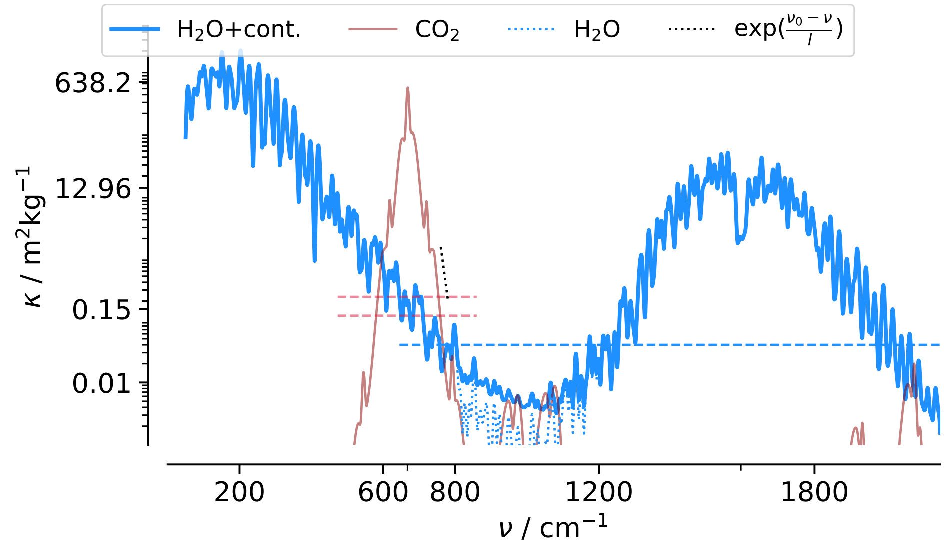

Figure 3. Mass absorption spectrum of H_2O (blue) and CO_2 (red) as a function of wave-number ν. Spectra are calculated at a wavenumber interval of $0.05\,\mathrm{cm}^{-1}$ for a temperature of $280\,\mathrm{K}$ and pressure of $850\,\mathrm{hPa}$ and smoothed by convolving with a Gaussian ($9\,\mathrm{cm}^{-1}$) filter to show the absorption envelope. For the dashed line envelope fit to the spectrum of CO_2. $l = 10.2\,\mathrm{cm}^{-1}$.

The implication of the above is that at wavenumbers ν and for W such that $\tau_{\nu,\mathrm{v}} \gg 1$, surface emission is attenuated and atmospheric emissions become constant. At wavelengths where $\tau_{\nu,\mathrm{v}} \gg 1$ and by virtue of only depending on T, changing the temperature of the atmosphere only changes the effective height of the emission, not the amount. This follows directly from Schwarzshild's equation (Chandrasekhar, 1960) for radiances, and can be extended to irradiances through the assumption of a constant diffusivity factor (Armstrong, 1968). For spectral regions where water vapor (or any absorber whose abundance is determined by T) controls the emission to space, these emissions become invariant of T_{sfc}. As T changes in physical space, the emitters simply redistribute themselves, retaining their abundances in temperature space, and emit the same amount. It is the conceptual heart of the fixed relative humidity assumption and provides the justification for **S3**, which can thus be formulated as

$$\delta F_\nu = \pi\delta\left(e^{-\tau_{\nu,\mathrm{v}}}\mathcal{B}_\nu(T_{\mathrm{sfc}})\right), \tag{9}$$

with $\mathcal{B}_\nu$ the Planck source function.

3.3 The support of the Planck response

Earth's globally averaged surface temperature is about $288\,\mathrm{K}$ and the average value of W is about $25\,\mathrm{kg\,m^{-2}}$. When considering the spectrum of $\kappa_{\nu,\mathrm{v}}$ (Fig. 3) it implies that $\forall\,\nu : \kappa_{\nu,\mathrm{v}} \gg W_{\mathrm{sfc}}^{-1}$, $\tau_{\nu,\mathrm{v}} \gg 1$. Modulo spectral smoothing hiding fine windows, this implies that the atmosphere is mostly opaque for $\nu < 800\,\mathrm{cm}^{-1}$ or $\nu > 1200\,\mathrm{cm}^{-1}$. The region in between, where $800\,\mathrm{cm}^{-1} < \nu < 1200\,\mathrm{cm}^{-1}$ is thus known as the atmospheric window. It defines the 'support' for changes in emission with warming.

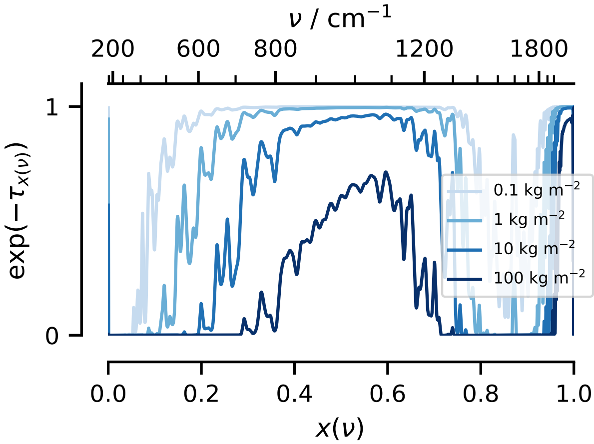

Figure 4. Spectral transmissivity plotted versus the cumulative black-body emission sensitivity, $x = (4\sigma T^3)^{-1} \int_0^{\nu} \left(\frac{\mathrm{d}\mathcal{B}_{\nu'}}{\mathrm{d}T}\right) \mathrm{d}\nu'$. The corresponding wave-numbers are indicated along the upper scale.

With the simplifications (**S1–S3**) the support, s, for a given $\tau_\nu(T) = \kappa_{\nu,\mathrm{v}} W(T)$ can be quantified as

$$s(T) = \frac{1}{4\sigma T^3} \int_0^{\infty} e^{-\tau_\nu(T)} \left(\frac{\mathrm{d}\mathcal{B}_\nu}{\mathrm{d}T}\right) \mathrm{d}\nu. \tag{10}$$

This defines the support s as the spectral transmissivity ($e^{-\tau_\nu(T)}$) weighted by the emission sensitivity to temperature ($\pi \mathrm{d}\mathcal{B}_\nu/\mathrm{d}T$) normalized by the black-body sensitivity. It is monotonic in W and defines the attenutaion of the spectral-response, going to zero as W becomes large, and unity as W vanishes. s differs from the broad-band transmissivity, as it describes the transparency of the atmosphere to changes in Planckian emissions, rather than the emissions them self.

The closing of the window with increasing W is illustrated in Fig. 4. Here $e^{-\kappa_\nu W(T)}$ has been plotted versus the fractional emission x, which is related to ν as

$$\mathrm{d}x = \frac{1}{4\sigma T^3} \left(\frac{\mathrm{d}\mathcal{B}_\nu}{\mathrm{d}T}\right) \mathrm{d}\nu \tag{11}$$

Choosing the x-axis in this manner stretches the ν axis so that equally spaced x intervals carry equal amounts of the radiative response. In terms of x, $s(T) = \int e^{-\kappa_\nu W(T)} \mathrm{d}x < 1$ is just the area under the curves in Fig. 4, which defines an effective interval of x over which an emission response to changing temperatures is 'supported', hence the name.

4 Spectral masking and climate sensitivity

The support, s, quantifies the idea of spectral masking. Jeevanjee et al. (2021a) refer to this as the cancellation of the spectral feedbacks. We find the idea of spectral masking more intuitive as it articulates the idea that only part of the spectrum participates in the radiative response to warming. Spectral masking can also be used more generally to understand radiative forcing from

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

long-lived greenhouse gases, which are the two ingredients needed to understand the climate sensitivity of a Simpsonian atmosphere whose albedo does not change with T.

4.1 Radiative response to warming

The radiative response parameter, λ, linearly relates changes in terrestrial emissions to space to changes in surface temperatures:

$$\Delta F = \lambda \delta T_{\mathrm{sfc}} \tag{12}$$

It is sometimes defined as a feedback parameter, and denoted by the same symbol, to quantify how much the system must warm to balance a given radiative forcing, but then adopts a different sign convention.

On the basis of **S1–S3** the radiative response of a cloud-free Simpsonian atmosphere to the change in the temperature, T, of an underlying black-body may be written as

$$\Lambda(T) = \pi \int e^{-\tau_{\nu,\mathrm{v}}} \left(\frac{\mathrm{d}\mathcal{B}_\nu}{\mathrm{d}T} \right) \mathrm{d}\nu = s(T)\, 4\sigma T^3. \tag{13}$$

We refer to Λ as the Simpsonian response. Applied to Earth, this leads to the expectation that for a cloud-free atmosphere

$$\lambda \approx \Lambda(T_{\mathrm{sfc}}) = s(T_{\mathrm{sfc}})\, 4\sigma T_{\mathrm{sfc}}^3, \tag{14}$$

a quantity that is sometimes denoted as the clear-sky radiative response, λ_{cs}.

Fig. 5 shows how s varies with T, how this influences $\Lambda(T)$, and how both depend on the choice of $W(T)$. To illustrate this dependence, three curves are shown. The black solid line evaluates s assuming $W = W_{\mathrm{sfc}}$, the teal curves are based on the assumption that $W = W_{\mathcal{R}}$. The contribution of continuum absorption to Λ is quantified by the dotted line which plots s and Λ for $W = W_{\mathcal{R}}$ in it absence.

The main findings are two fold. First, and qualitatively, Fig. 5 demonstrates that for a fixed $\mathcal{R} \neq 0$ atmosphere, Λ maximizes at a value near $2.7\,\mathrm{W\,m^{-2}\,K^{-1}}$ at a temperature, $T_{\max} \approx 260\,\mathrm{K}$. Hence, roughly speaking, in the ice phase ($T < T_{\max}$), $\mathrm{d}\Lambda/\mathrm{d}T \approx 0.02\,\mathrm{W\,m^{-2}\,K^{-2}}$, while for a surface temperature supporting a liquid phase, $\mathrm{d}\Lambda/\mathrm{d}T \approx -0.1\,\mathrm{W\,m^{-2}\,K^{-2}}$.

Second, and quantitatively, the simple estimates of $\lambda_{\mathrm{cs}} \approx \Lambda(T_{\mathrm{sfc}})$ are surprisingly accurate. For instance, for a global mean surface temperature of about $290\,\mathrm{K}$ full radiative transfer calculations for an atmosphere in radiative convective equilibrium yield an estimate of $\lambda_{\mathrm{cs}} = 2.2\,\mathrm{W\,m^{-2}\,K^{-1}}$, in excellent agreement with more precise estimates, e.g., as in (McKim et al., 2021). While the quantitative fidelity may benefit to some extent from errors of different approximations to some degree compensating one another, these do not seem to be large effects. To the extent λ_{cs} can be usefully approximated by $\Lambda(T_{\mathrm{sfc}})$, it demonstrates that this response is something that, given knowledge of the water vapor absorption spectra, is quite easy to understand. Essentially the reduction in λ_{cs} from what would be expected from a blackbody, measures how effective water vapor is at controlling emission to space, and thereby masking the spectral response of emissions to surface warming.

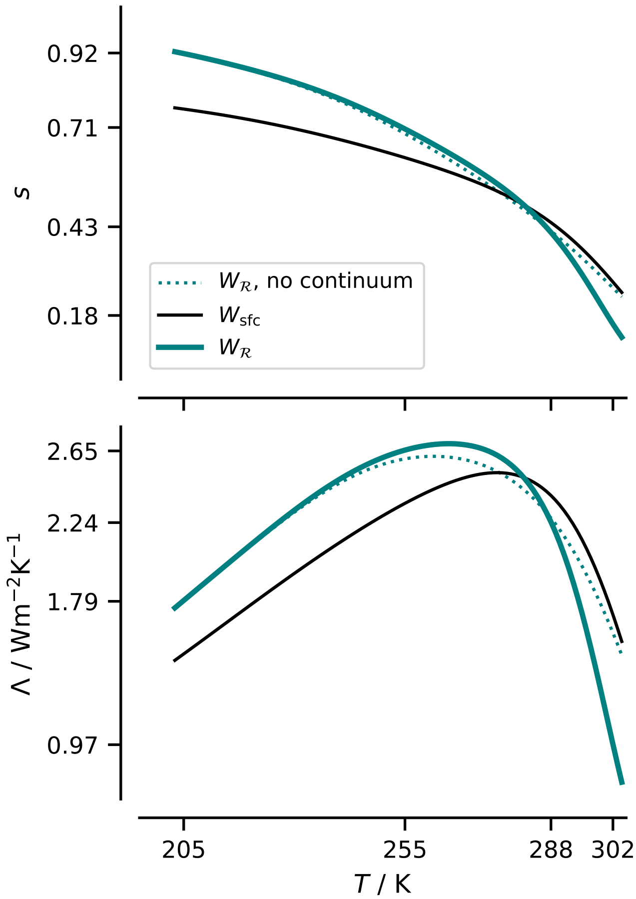

Figure 5. Variation of the support, $s(T)$, (upper) and the radiative response to warming, λ with T for different models of $W(T)$. Solid lines show calculations with the inclusion of continuum absorption the dotted line, for reference, shows the response in the absence of this absorption.

4.2 CO_2 radiative forcing and climate sensitivity

Similar ideas can be used to estimate the radiative forcing from a doubling of atmospheric CO_2, $F_{2\times}$, and hence the climate sensitivity. Here the CO_2 absorption feature around $667.5\,\mathrm{cm}^{-1}$ masks emission from the lower troposphere (or surface) and increasing CO_2 increases the masking. Unlike for the case of water, for which the masked emissions don't change with warming, increasing the spectral footprint of CO_2 masking through a doubling of its concentration, leads to more of the near surface emissions being supplanted by emissions from the tropopause.

Based on these arguments, which are similar to those developed by Wilson and Gea-Banacloche (2012); Jeevanjee et al. (2021b), the emission at the top of the clear-sky, water vapor free, atmosphere can be estimated as

$$F = \sigma T_{\rm cp}^4 - \pi \int e^{-\tau_{\nu,\rm c}} \left[\mathcal{B}_\nu(T_{\rm cp}) - \mathcal{B}_\nu(T_{\rm sfc})\right] \mathrm{d}\nu, \tag{15}$$

https://doi.org/10.5194/egusphere-2022-1460
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with $\tau_{\nu,\mathrm{c}}$ the CO_2 optical depth at wavenumber ν, which with **S1** becomes

$$\tau_{\nu,\mathrm{c}} \approx \kappa_{\nu,\mathrm{c}} C \quad \text{with} \quad C = \int_0^\infty \rho_\mathrm{c}\,\mathrm{d}z. \tag{16}$$

This defines C as the CO_2 burden, i.e., the analogy to W for water vapor. Physically, the first two terms in Eq. (15) describe emissions from the tropopause region, while the third term describes the emissions from the surface which are not masked by CO_2. The forcing from a doubling of atmospheric CO_2 can then be calculated as

$$F_{2\times} = \pi \int \left(e^{-\tau_{\nu,\mathrm{c}}} - e^{-2\tau_{\nu,\mathrm{c}}}\right)\left[\mathcal{B}_\nu(T_\mathrm{sfc}) - \mathcal{B}_\nu(T_\mathrm{cp})\right]\mathrm{d}\nu, \tag{17}$$

It shows that in the absence effects of water vapor, $F_{2\times}$ depends on the absorption spectrum of CO_2 and the temperature of the surface and at the tropopause, which we take to be at the cold-point temperature.

Quantitatively we can evaluate Eq. (17) across a range of surface temperatures, and for different tropopause temperatures. For Earth, with an average temperature of $288\,\mathrm{K}$ and a coldpoint temperature of $190\,\mathrm{K}$ this implies $F_{2\times} = 5.4\,\mathrm{W\,m^{-2}}$. For a $13\,\mathrm{K}$ colder surface the forcing reduces by $1\,\mathrm{W\,m^{-2}}$, or approximately $0.075\,\mathrm{W\,m^{-2}\,K^{-1}}$, thereby increasing at rate greater than is compensated by the cold-state increase in λ with warming. In addition with T_cp $20\,\mathrm{K}$ warmer, as might be more representative of the extra tropics (Tegtmeier et al., 2020), $F_{2\times}$ is, for the same surface temperature, approximately $0.7\,\mathrm{W\,m^{-2}}$ smaller – all of which ensures that the sensitivity ($F_{2\times}/\lambda$) is a strictly increasing function of temperature, increasingly so as λ starts decreasing for $T > 270\,\mathrm{K}$ (Fig. 5).

One challenge for the above estimate of $F_{2\times}$ is that it does not account for the effect of water vapor in masking either emissions from the surface, or from CO_2. The former can roughly be estimated, from Fig. 3 by noting that the wings of the absorption envelope of the $667.5\,\mathrm{cm^{-1}}$ CO_2 line corresponds to a water vapor absorption of about $0.15\,\mathrm{m^2\,kg^{-1}}$, which becomes optically thick for $W = 6.67\,\mathrm{kg\,m^{-2}}$. Following Eq. (6), this value of W corresponds to a temperature of about $275\,\mathrm{K}$, which effectively limits

$$\begin{aligned} F_{2\times} = \pi \int & \left(e^{-\tau_{\nu,\mathrm{c}}} - e^{-2\tau_{\nu,\mathrm{c}}}\right) \\ & \left[\mathcal{B}_\nu(\min(275, T_\mathrm{sfc})) - \mathcal{B}_\nu(T_\mathrm{cp})\right]\mathrm{d}\nu, \end{aligned} \tag{18}$$

so that for T_cp ranging from $194\,\mathrm{K}$ to $204\,\mathrm{K}$, $F_{2\times}$ saturates at $4.3\,\mathrm{W\,m^{-2}}$ to $3.9\,\mathrm{W\,m^{-2}}$. These overlap effects are explored in greater detail and quantified by Jeevanjee et al. (2021b), which agree well with the above simple estimates and rationalizes empirical fits to more complete calculations as in Huang et al. (2016).

Combining this estimate of $F_{2\times}$ with the earlier estimate of λ_cs yields an estimate for the clear-sky climate sensitivity,

$$\mathcal{S}_\mathrm{cs} \equiv \frac{F_{2\times}}{\lambda_\mathrm{cs}} \approx \frac{4.3}{2.2} \approx 2\,\mathrm{K}, \tag{19}$$

as compared to $2.1\,\mathrm{K}$ derived from radiative convective equilibrium calculations using full line-by-line radiative transfer.

Consistent with this difference, but not accounted for in these calculations, CO_2 reclaims some of the spectrum from H_2O, thereby increasing s and hence increasing $\mathcal{S}$ (Kluft et al., 2021; Seeley and Jeevanjee, 2021). This illustrates how the overlap

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

between H_2O and CO_2 reduces the sensitivity of the atmosphere to increasing CO_2, both by reducing the forcing and by increasing the radiative response.

4.3 Simple expressions for forcing, feedback and Earth's clear-sky climate sensitivity.

The above analysis results in the following estimate for the clear-sky climate sensitivity:

$$\mathcal{S}_{\mathrm{cs}} \approx \frac{F_{2\times}}{\Lambda(T_{\mathrm{sfc}})}, \tag{20}$$

with $F_{2\times}$ given by Eq. (18) and Λ by Eq.(13).

The physical content of Eq. (20) can be better illustrated by some simple approximations to the envelope of the absorption spectrum of water vapor and CO_2, and its ability to mask spectral emission. For instance, By approximating $e^{-\tau_{\nu,\mathrm{v}}}$ by unity outside, and zero inside, the atmospheric window, which based on Fig. 4 we take to be between $800\,\mathrm{cm}^{-1}$ to $1200\,\mathrm{cm}^{-1}$, the Simpsonian response can be expressed simply as the integral of the response over the window,

$$\Lambda \approx \pi \int\limits_{800}^{1200} \left(\frac{\mathrm{d}\mathcal{B}_\nu}{\mathrm{d}T}\right) \mathrm{d}\nu. \tag{21}$$

Contributions from all other wavenumbers are masked by the Simpsonian response of water vapor, see also Colman and Soden (2021) who recently came to a similar conclusion.

Likewise for CO_2, Fig. 3 shows that the envelope of the CO_2 absorption spectrum falls off exponentially as $\alpha e^{\|\nu-\nu_{\mathrm{c}}\|/l}$, with $l \approx 10.2\,\mathrm{cm}^{-1}$ as calculated by Jeevanjee et al. (2021b). Wilson and Gea-Banacloche (2012) adopted this as a parameterization of the CO_2 spectrum, which implies that for a CO_2 burden of C, $\tau_{\nu,\mathrm{c}} > 1$ for

$$\nu_{\mathrm{c}} - l\ln(\alpha C) < \nu < \nu_{\mathrm{c}} + l\ln(\alpha C). \tag{22}$$

From this it follows that for a burden of $2C$ the atmosphere becomes optically thick for the larger interval, larger by the amount $2l\ln(2)$. As the interval is small compared to the range of wavenumbers over which $\mathcal{B}_\nu$ varies, $\mathcal{B}_\nu(T) \approx \mathcal{B}_{\nu_{\mathrm{c}}}(T)$ over the interval. With these simplifications the expression for the forcing simplifies to

$$F_{2\times} \approx \pi 2l\ln(2)\left[\mathcal{B}_{\nu_{\mathrm{c}}}(\min(275, T_{\mathrm{sfc}})) - \mathcal{B}_{\nu_{\mathrm{c}}}(T_{\mathrm{cp}})\right] \tag{23}$$

and the clear-sky climate sensitivity becomes

$$\mathcal{S}_{\mathrm{cs}} \approx 2l\ln(2)\frac{\mathcal{B}_{\nu_{\mathrm{c}}}(\min(275, T_{\mathrm{sfc}})) - \mathcal{B}_{\nu_{\mathrm{c}}}(T_{\mathrm{cp}})}{\int_{800}^{1200}\left(\frac{\mathrm{d}\mathcal{B}_\nu}{\mathrm{d}T}\right)\mathrm{d}\nu} = 2.37\,\mathrm{K}. \tag{24}$$

This estimate, which is derived from rather simple reasoning, agrees well with calculation as given in Eq. (20), and with the detailed radiative transfer calculations by Kluft et al. (2019). The main purpose of Eq. (24), is less about the quantitative fidelity of the reasoning, and more to emphasize that the clear sky climate sensitivity, and its dependence on things like surface temperature, forcing amount, and tropopause temperature, is something we understand.

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

5 Inferences for Earth's atmosphere

The above ideas have, or are, begin developed simultaneously by other groups, as referenced above. Here we extend the scope of inquiry by exploring what they imply for the effects of clouds and for how temperature mediates the atmospheres radiative response to forcing (CO_2 changes) and to surface warming.

5.1 Clouds

Understanding of the clear-sky climate sensitivity provides the foundation for thinking about how clouds change the picture. Here the idea of masking is again useful. What differentiates clouds from water vapor is that they are neither colorful, nor Simpsonian. The first means that clouds will mask the forcing and the radiative response to warming, weakening both. The fact that clouds are neither colorful, nor Simpsonian, raises the possibility that if clouds change their temperature with surface warming, they may unmask the spectral response, and perhaps even enhance the radiative response to forcing.

The effect of clouds on the radiative forcing can be understood as follows. If clouds are located high enough in the troposphere, they will mask the increased masking of surface emissions from increases in atmospheric CO_2, or in the case they are lower in the troposphere, set a colder baseline emission (effective value of T_{sfc} in Eq. 17) being masked by increased CO_2. In both cases the radiative forcing will be reduced. Assuming the effect of high-clouds is dominant, an optically thick high-cloud fraction, $f_{\mathrm{h}} = 0.25$, would reduce the forcing by a commensurate amount, reducing $F_{2\times}$, from $4.9\,\mathrm{W\,m^{-2}}$ as calculated by Kluft et al. (2019) to $3.7\,\mathrm{W\,m^{-2}}$. As a comparison, Myhre et al. (1998) estimate a slightly larger $27\,\%$ reduction in CO_2 forcing due to clouds.

When the cloud top emission temperature, T_{cld} does not change with warming, clouds also mask window emissions in proportion to their (optically thick) cloud fraction, an effect which was also developed conceptually by McKim et al. (2021). Given estimates of total cloud fraction, f_{t}, of about 0.6, this implies a sharp reduction in the ability of the atmosphere to radiatively respond to warming, reducing λ from $2.2\,\mathrm{W\,m^{-2}\,K}$ to $0.9\,\mathrm{W\,m^{-2}\,K^{-1}}$. Because all clouds contribute to the masking of emissions within the window, this is a markedly stronger effect than the reduction of the forcing, and increasing $\mathcal{S}$ by $(1-f_{\mathrm{h}})/(1-f_{\mathrm{t}}) \approx 1.875$. Behaving in this manner clouds can be expected to considerably amplify Earth's equilibrium climate sensitivity, increasing it to about $4\,\mathrm{K}$.

To the extent clouds also increase their temperature with warming, they will unmask parts of the spectrum in which emissions changes would otherwise be controlled, and hence masked, by water vapor. This can support a stronger radiative response. Consider the case of trade-wind clouds topping a warm marine boundary layer, with $T_{\mathrm{cld}} \approx T_{\mathrm{sfc}} - 15\,\mathrm{K}$. Furthermore, suppose, that due to their strong coupling with the surface, $\delta T_{\mathrm{cld}} \approx \delta T_{\mathrm{sfc}}$. By virtue of topping the near surface concentration of moisture that would otherwise control the radiation, clouds reclaim spectral territory and actually increase the radiative response to warming. From Fig. 5, clouds with tops at $288\,\mathrm{K}$ will radiate twice as much energy per degree of warming than would the surface at $305\,\mathrm{K}$. This suggests that shallow boundary layer clouds, even small ones that cover most of the tropical oceans but generally go unnoticed (Mieslinger et al., 2022; Konsta et al., 2022), might play a role in stabilizing the climate. In contrast,

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

over the cold extra-tropics, where the radiative response to warming increases with temperature, high clouds will be less effective in making use of any spectral emissions that they reclaim from water vapor.

To make these ideas more precise we can model the effects of clouds on the radiative response to warming such that

$$\lambda \approx (1-f)\,\Lambda_{\mathrm{sfc}} + f\left(\frac{\delta T_{\mathrm{cld}}}{\delta T_{\mathrm{sfc}}}\right)\Lambda_{\mathrm{cld}}, \tag{25}$$

where we introduce Λ_{x} to denote $\Lambda(T_{\mathrm{x}})$, for notational convenience. The first term in Eq. (25), describes the masking of the clear-sky response (assuming $\lambda_c s \approx \Lambda_{\mathrm{sfc}}$) by clouds, as in McKim et al. (2021), the second term describes the emission response across the spectrum as reclaimed by clouds.

Eq. (25) demonstrates how, in the warm tropical atmosphere, where precipitating convection is embedded in a nearly saturated atmosphere (Bretherton and Peters, 2004), clouds may be especially important for the radiative response to warming. As the window closes, $\lambda_{\mathrm{cs}} \to 0$, so that there is nothing left for clouds to mask, which is to say that the first term in Eq. (25) vanishes independent of f. In this case clouds with cold cloud tops will carry the entire radiative response, and its magnitude will be in proportion to the cloud fraction and the cloud top temperature change. This would provide a thermostat for the tropical hothouse, one whose effectiveness will depend on the degree to which cloud top temperature changes are constrained by the radiative cooling in the clear-sky atmosphere, which is still a matter of some debate (Zelinka and Hartmann, 2010, 2011; Bony et al., 2016; Seeley et al., 2019).

For the case when the window is still open ($\Lambda_{\mathrm{sfc}} \neq 0$), and cloud masking limits the radiative response, the net effect of clouds can be more easily seen by rearranging Eq. (25) as follows:

$$\lambda \approx \Lambda_{\mathrm{sfc}}\left[1 + f\left(\frac{\delta T_{\mathrm{cld}}}{\delta T_{\mathrm{sfc}}}\eta - 1\right)\right], \quad \text{with} \quad \eta = \Lambda_{\mathrm{cld}}/\Lambda_{\mathrm{sfc}}. \tag{26}$$

For T_{sfc} less than the value of T at which Λ maximizes, which from Fig. 5 is about $260\,\mathrm{K}$, $\eta < 1$ requires clouds to warm more than the surface to offset their masking of window emissions. Over a warm surface, even in the dry tropics, $\eta > 1$, with $\eta \approx 3$ even possible for mid-level clouds above a very warm surface. In this case, and if $\eta\delta T_{\mathrm{cld}} > \delta T_{\mathrm{sfc}}$, clouds will also increase the radiative response to warming in proportion to their coverage.

This analysis can be generalized to clouds distributed over multiple layers, by working ones way down through the successive contribution of layers of non-overlapped clouds:

$$\lambda = \Lambda_{\mathrm{sfc}}\left[1 + \sum_i f'_{\mathrm{i}}\left(\frac{\delta T_{\mathrm{cld},i}}{\delta T_{\mathrm{sfc}}}\eta_i - 1\right)\right] \tag{27}$$

where f'_i denotes the cloud fraction for layer i (increasing downward) that is not geographically masked by clouds at layers $j < i$.

This analysis identifies ways in which the amount and distribution of clouds can play an important role in setting the radiative response to warming, even if cloud coverage and temperatures do not change, and how, depending on how cloud top temperatures changes, clouds can either increase or decrease the clear-sky climate sensitivity. This analysis also demonstrates that the role clouds plays can be quite different in the cold extra-tropics versus the warm tropics.

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

5.2 Latitudinal variations

The above considerations raise the question as to how the covariability of clouds and surface temperature might influence Earth's radiative response geographically. For instance, from a purely radiative point of view, the idea that the poles should warm disproportionately is a curious one, as the radiative forcing from a doubling of atmospheric CO_2 is proportional to $T_{\mathrm{sfc}} - T_{\mathrm{cp}}$, which is much smaller in the polar regions, and the radiative response to warming is, by virtue of the absence of water vapor to mask surface emissions, particularly large. As a consequence of these effects, the climate sensitivity is expected to increase with temperature (as the reduced sensitivity at very cold temperatures is compensated by reduced forcing), more strongly so as $s(T)$ begins to decrease with T for $T > 270\,\mathrm{K}$. As the window closes this leads to the expectation of surface temperatures decoupling from the radiative response, and runaway warming, something that recent work (Kluft et al., 2021; Seeley and Jeevanjee, 2021) has shown is arrested by the presence of long-lived greenhouse gases anchoring part of the spectral response to surface temperatures.

Putting aside the warming catastrophe, which becomes important at temperatures and humidities not generally found on the present day earth, we investigate how surface the geographic distributions of surface temperature and cloudiness influence Earth's sensitivity to forcing locally. In Fig. 6, we compare estimates of the S with λ estimated following Eq. (25) assuming different cloud behaviors and with T_{sfc}, and $W_{\mathrm{sfc}}(T_{\mathrm{sfc}})$ (rather than $W_{\mathcal{R}}$) used to calculate Λ_{sfc} and with $W_{\mathcal{R}}$ used to calculate Λ_{cld}. This is an admittedly crude way to treat the variation of W with height at different geographic regions, but using W_{sfc} for the cloud term as well does not change the answer appreciably. Without performing the full radiative transfer with a realistic distribution of clouds and water vapor, it is also not obvious how to calculate the degree to which clouds mask the forcing, so again a very simple approach is taken, in line with the efforts of this paper to develop conceptual understanding. The masking fraction, $f_{2\times}$ is assumed to come from a single layer of clouds whose coverage is, ad hoc, based on the cloud-top temperature as

$$f_{2\times} = \alpha \left[\frac{T_{\mathrm{sfc}} - T_{\mathrm{cld}}}{T_{\mathrm{sfc}} - T_{\mathrm{cp}}} \right] f_{\mathrm{t}}. \tag{28}$$

with $\alpha = 1.9$ a tuning constant chosen so that $\overline{F}_{2\times}$ matches the estimate of $3.7\,\mathrm{W\,m^{-2}}$ of more detailed calculations.

To calculate the effects of clouds on λ, and hence estimate the cliamte sensitivity, $\mathcal{S}$, three bounding cases are considered. One wherein clouds are made transparent, both for the forcing and the radiative response to warming. A second where $\delta T_{\mathrm{cld}} \approx \delta T_{\mathrm{sfc}}$, and a third where $\delta T_{\mathrm{cld}} = 0$. Because$\mathcal{S}$ is defined as a global (or statistical) quantity, it is estimated as $\overline{F_{2\times}}/\overline{\Lambda}$.

For the case of transparent clouds, $f = 0$, values of $F_{2\times}/\Lambda$ vary latitudinally from a low value ($0.7\,\mathrm{K}$) over the South Pole, to a high value ($3.4\,\mathrm{K}$) over the ITCZ region just north of the Equator. $\mathcal{S} = 2.4\,\mathrm{K}$ is similar to the clear-sky estimates obtained previously, using global mean quantities. The case when $\delta T_{\mathrm{cld}} \approx \delta T_{\mathrm{sfc}}$ leads to $\mathcal{S} = 1.7\,\mathrm{K}$ with reductions most pronounced in the tropics where additional emissions from clouds occurs in an atmosphere that is less masked by water vapor. Given the idea that high-clouds maintain a fixed temperature, this case might seem extreme, then again, warming along the moist adiabat is upward amplified, so that the case of fixed cloud height actually implies $\delta T_{\mathrm{cld}} > \delta T_{\mathrm{sfc}}$, which can be thought of as a form of lapse-rate feedback. For the third ($\delta T_{\mathrm{cld}} = 0$) case, clouds mask the radiative response, and $\mathcal{S}$ increases considerably, inverting

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

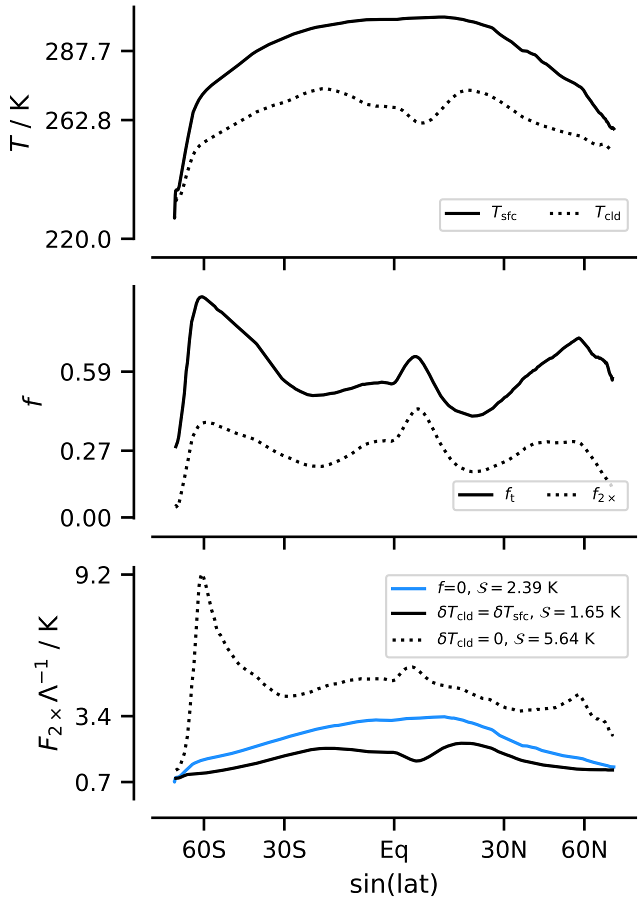

Figure 6. Latitudinal distribution of T_{sfc} and T_{cld} (upper), total cloud fraction f_{t} and fraction assumed to mask CO_2 forcing, $f_{2\times}$ (middle); and the ratio of the forcing $F_{2\times}$ to the radiative response to warming, λ for different assumptions about clouds (lower).

its geographic structure to be more poleward amplified. This behavior depends on $f_{2\mathrm{x}}$, as for $f_{2\mathrm{x}} = f$, the masking of the forcing cancels the masking of the response and $\mathcal{S}$ follows the transparent case.

5.3 Climate sensitivity

To answer the question as to whether clouds increase or decrease Earth's equilbrium climate sensitivity we first ask how much clouds have to warm to compensate their masking effects. By adjusting δT_{cld} until $\mathcal{S} = \mathcal{S}_{\mathrm{cs}}$ (which corresponds to the transparent, $f = 0$, estimate in Fig. 6) we find that for $\delta T_{\mathrm{cld}} \approx {}^{1}\!/_{2}\,\delta T_{\mathrm{sfc}}$ the warming of clouds compensates their masking. Given that clouds also mask surface albedo changes with warming, whose assessed value of $0.35\,\mathrm{W\,m^{-2}\,K^{-1}}$ (Forster et al., 2021) is believed to be half of what they would be in a cloud-free atmosphere (Pistone et al., 2014), clouds only have to warm by about ${}^{1}\!/_{4}\,\delta T_{\mathrm{sfc}}$ to start having a net cooling effect relative to that expected from a clear sky, similar to what Kluft et al. (2019)

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

estimate for the warming of high clouds in the tropics. Hence unless the amount of clouds change it seems that they make the system less, rather than more, sensitive to forcing. This assertion is at odds with literature that does not account for the ability of clouds to unmask the spectral response to warming. Even if their coverage does change, the assessed feedback from cloud amount changes is only $0.2\,\mathrm{W\,m^{-2}\,K^{-1}}$ (Forster et al., 2021), with recent work suggesting that it may be even smaller (Myers et al., 2021; Vogel et al., 2022). To balance this reduction in the radiative response would require clouds to warm on average by $\delta T_{\mathrm{cld}} \approx 0.4\,\delta T_{\mathrm{sfc}}$. Based on these estimates, a null hypothesis of no net cloud contribution to warming, but with a residual surface albedo feedback of $0.35\,\mathrm{W\,m^{-2}\,K^{-1}}$, doesn't seem far fetched – all the more so because it yields an estimate of $\mathcal{S}$ of $2.96\,\mathrm{K}$.

While this estimate does not fundamentally change our view on the value of climate sensitivity, it does change how we think about the factors that influence it, and thereby suggests a different research programme for narrowing uncertainty in its estimation. Instead of quantifying the feedbacks (planck, water vapor, lapse-rate) that emerge from conceptualizing the atmosphere as grey, such a programme would:

1. quantify $\mathcal{S}_{\mathrm{cs}}$ as the clear-sky Simpsonian response to warming;

2. quantify 'corrections' from (i) cloud masking; (ii) cloud changes; (iii) non-Simpsonian water vapor changes.

Calculating $\mathcal{S}_{\mathrm{cs}}$ based on the clear-sky Simpsonian response, and distinguishing it from $\mathcal{S}$, is justified by the strong theoretical foundation for the Simpsonian response, and an ability to estimate it from existing measurements with quantifiable uncertainty. For this reason, and because there is no a priori expectation for the sign of the corrections, $\mathcal{S}$ may then by estimated by taking $\mathcal{S}_{\mathrm{cs}}$ as a prior subject to Bayesian updates by the 'corrections'.

6 Conclusions

Assuming that the relative humidity, $\mathcal{R}$, does not vary with temperature, T, places a strong constraint on Earth's radiative response to warming (Simpson, 1928; Nakajima et al., 1992; Ingram, 2010). In this limit, the radiative response to warming in those spectral intervals where water vapor is optically thick is nullified – something we call spectral masking. By accounting for this spectral masking it becomes straightforward to derive an expression for Earth's clear-sky radiative response to warming. This expression, given by Eq. (13), is quantitatively accurate and physically informative.

The concept of spectral masking can also be extended to long-lived trace gases, which allows for simple expressions describing their radiative response to changes in their concentration, i.e., radiative forcing. With some further, more heuristic, simplifications analytic expressions can be derived to describe the radiative forcing from changing concentrations of CO_2, which when combined with the radiative response to warming yields an expression, Eq. (20), for the clear-sky equilibrium climate sensitivity, $\mathcal{S}_{\mathrm{cs}}$. These ideas help understand and quantify state dependence, i.e., $\mathcal{S}_{\mathrm{cs}}$ depending on temperature (Caballero and Huber, 2013; Bloch-Johnson et al., 2021), predicting that it increases with temperature, increasingly so for $T_{\mathrm{sfc}} > 270\,\mathrm{K}$.

Spectral masking also provides a basis for thinking about how clouds $\mathcal{S}_{\mathrm{cs}}$. Clouds – even for no change in geographic coverage – can both mask emissions in parts of the spectrum not controlled by selective absorbers, namely the atmospheric

https://doi.org/10.5194/egusphere-2022-1460
Preprint. Discussion started: 2 January 2023

window for the case of water vapor, and reclaim spectral territory by increasing the emission of radiation to space with warming, across the spectrum. Moreover, by virtue of locating at a different, usually colder, temperature than the surface, clouds that warm with the surface, amplify the radiative response over a warm surface, and damp the response over a cold surface. Clouds introduce an additional state dependence to the climate sensitivity, one that depends on their covariability with T and T_{sfc}. This state dependence renders estimates of climate sensitivity sensitive to not just how clouds change, but also their base-state distribution. It also means that Earth's geographic tendency to have more clouds where it is colder moderates geographic variations in the ratio of the local radiative forcing to the local radiative response, $F_{2\times}/\lambda$, and is thus a source of the poleward amplification of warming.

This analysis also illustrates some surprising properties of clouds: (i) the ability of diminutive clouds in the tropics, whose cloud top temperatures are more closely bound to surface temperature changes, to substantially increase the radiative response of the tropical atmosphere to warming; (ii) the importance of even small temperature changes in regions of deep convection for amplifying the radiative response of the moist tropics to warming; (iii) the importance of cloud masking at high-latitudes for increasing the sensitivity of regions whose clear-sky atmosphere would otherwise not be expected to be particularly susceptible to forcing; and (iv) the many, but poorly appreciated, ways in which clouds reduce the climate sensitivity. Small changes in cloud-top temperatures, or in the amount of very thin low clouds atop the tropical boundary layer can compensate or compound changes in optically thick clouds. This renders the *net* cloud contribution to warming ambiguous, and adds weight to the value of a theoretical understanding of the clear-sky climate sensitivity and the components which contribute to it.

The revised conceptual framework, combined with estimates of surface albedo feedbacks from the literature, allows us to quantify Earth's equilibrium climate sensitivity. The result, 3 K, doesn't meaningfully differ from values proposed by recent assessments adopting different approaches. Its distinguishing feature is the robustness and transparency of its reasoning. Robustness builds confidence in the value; transparency articulates new research programmes that might help narrow uncertainty further. For instance, recent assessments (Stevens et al., 2016; Sherwood et al., 2020) take a state of no knowledge of the radiative response to warming as a prior for Bayesian estimates. The arguments developed here, and independently by other authors, demonstrate that we quantitatively understand the clear-sky sensitivity, and that there is not an a priori reason to think that clouds should change this in one way or another. Hence it seems reasonable to adopt the $\mathcal{R}(T)$, or Simpsoninan, clear-sky sensitivity as a prior for the overall climate sensitivity $\mathcal{S}$. Moreover, our calculations show how observations, coupled with more precise radiative transfer calculations, are probably sufficient to determine its value with quantifiable uncertainty. Starting from the clear sky, and adding corrections arising from: (i) estimates of how $\mathcal{R}$ might change; (ii) masking by clouds; and (iii) cloud changes with warming, could be used as a part of a Bayesian updating to better quantify $\mathcal{S}$.

This study emphasizes how the role of clouds in modulating the climate sensitivity of a planet with fixed albedo is determined by their temperature, how this temperature differs from the temperature of the surface, and how it changes. Observations in the most transparent parts of the spectrum, can quantify these values and thus stand to advance understanding the most, in particular to better understand the role diminutive clouds whose temperatures are coupled to the surface, in amplifying the radiative response to warming, and of high-clouds in cold regions, in damping it. Aligning the analysis of more complex models with the physics of the problem, e.g., by evaluating cloud responses in temperature and vapor, rather than in geographic space, offers opportunities for gleaning more insight as to the plausibility of the processes these models simulate, or parameterize, and their ultimate role in modifying Earth's clear-sky climate sensitivity.

Code availability. The code used to produce alll figurs and make all calculations is provides as a Python notebook on Zenodo

Author contributions. The presented concepts and ideas have been developed by BS and LK during a joint lecture. BS has performed the analysis, created the figures, and written the original draft. BS and LK have reviewed and edited the submitted manuscript.

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