Recent improvements in the understanding of water vapour feedback

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Outlook

- Background
- Water vapour feedbacks in idealized cases :The “Simpson law”
- Clear-sky water vapour feedbacks in realistic cases
- Implication for the clear-sky lapse-rate feedback
- All sky multi-model results
- Conclusion

First estimate of the water vapour feedback

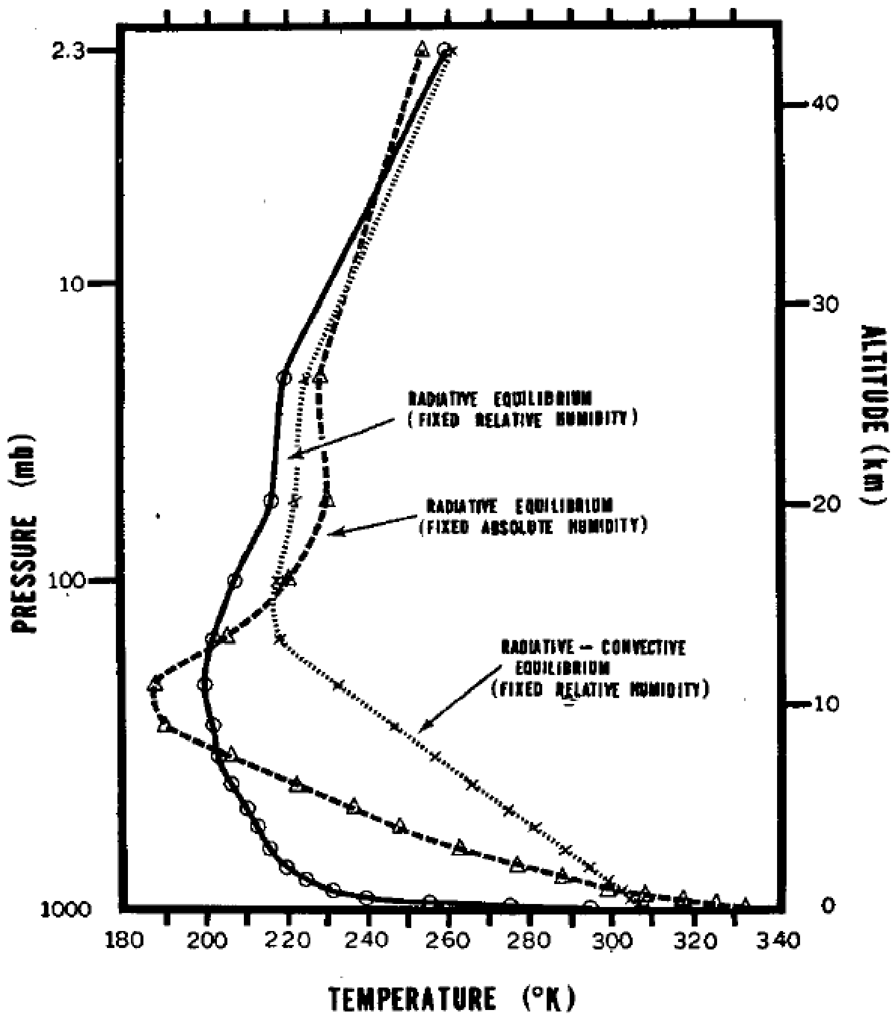


FIG. 5. Solid line, radiative equilibrium of the clear atmosphere with the given distribution of relative humidity; dashed line, radiative equilibrium of the clear atmosphere with the given distribution of absolute humidity; dotted line, radiative convective equilibrium of the atmosphere with the given distribution of relative humidity.

[Manabe & Wetherald, 1967]

TABLE 4. Equilibrium temperature of the earth's surface (°K) and the CO_2 content of the atmosphere.

	Average cloudiness		Clear	
CO_2 content (ppm)	Fixed absolute humidity	Fixed relative humidity	Fixed absolute humidity	Fixed relative humidity
150	289.80	286.11	298.75	304.40
300	291.05	288.39	300.05	307.20
600	292.38	290.75	301.41	310.12

2) Generally speaking, the sensitivity of the surface equilibrium temperature upon the change of various factors such as solar constant, cloudiness, surface albedo, and CO_2 content are almost twice as much for the atmosphere with a given distribution of relative humidity as for that with a given distribution of absolute humidity.

Radiative forcing-feedback (or response) framework

The radiative forcing ΔQ is the **change in the net radiative flux** (in $W.m^{-2}$) at the top of atmosphere due to a change in an external forcing (a driver of climate change) **before surface temperature adjusts** to this perturbation

The "climate feedback parameter" λ is the **sensitivity of the net radiative flux** at the top of atmosphere **to a change in the global mean surface temperature T_s** (in $W.m^{-2}.K^{-1}$)

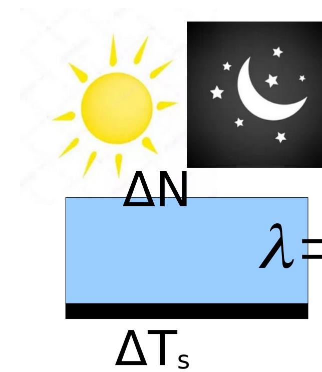
$$\Delta N = \Delta Q + \lambda \Delta T_s$$

Change in net flux at the TOA

radiative forcing

"climate feedback parameter"

Change in global mean **surface** temperature



$$\lambda = \frac{dN}{dT_s}$$

Here $\lambda < 0$. The opposite sign convention is also used

The global mean surface **temperature change at equilibrium:** $\Delta T_s^e = -\frac{\Delta Q}{\lambda}$

Climate feedbacks

Classical decomposition (***specific humidity***)

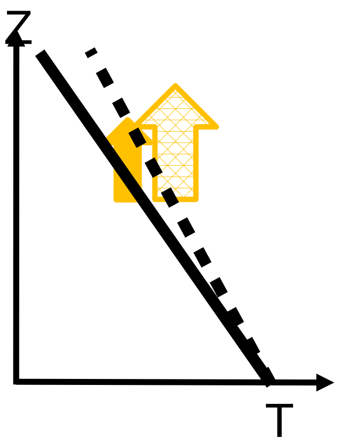
$$\lambda = \lambda_T + \lambda_W + \lambda_C + \lambda_\alpha$$

temperature — water vapor — clouds — surface albedo

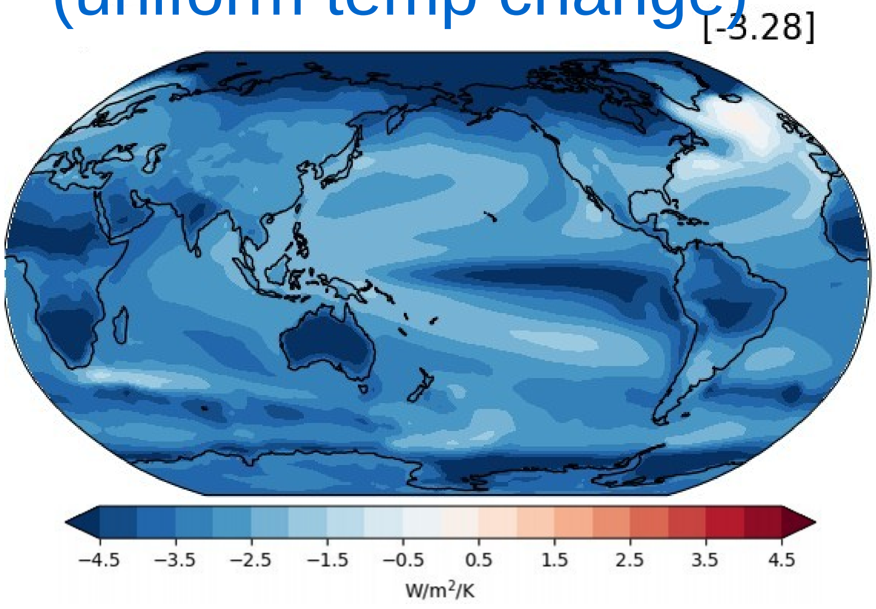
$$\lambda_T = \lambda_P + \lambda_L$$

Planck → Uniform vertical temperature change

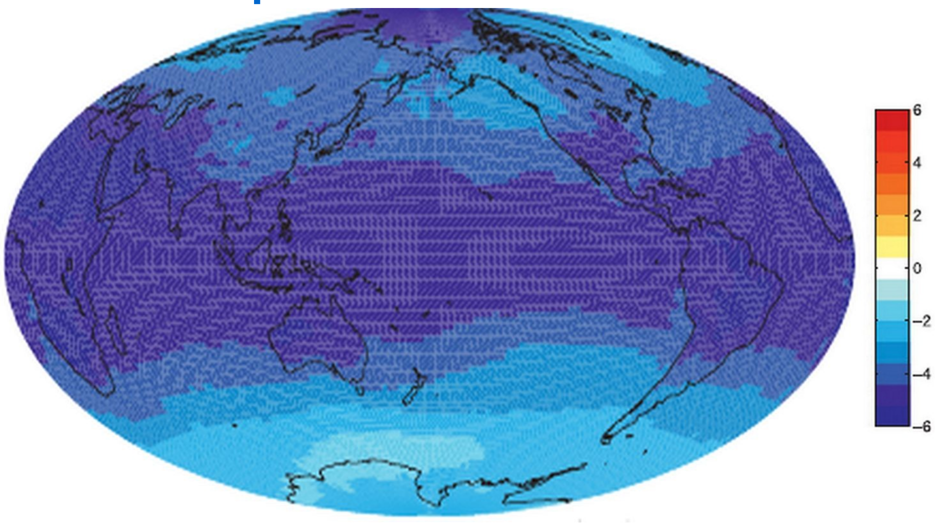
lapse rate → Departure from uniform vertical temperature change



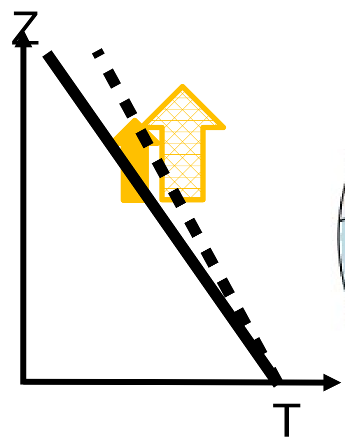
Planck feedback (uniform temp change)

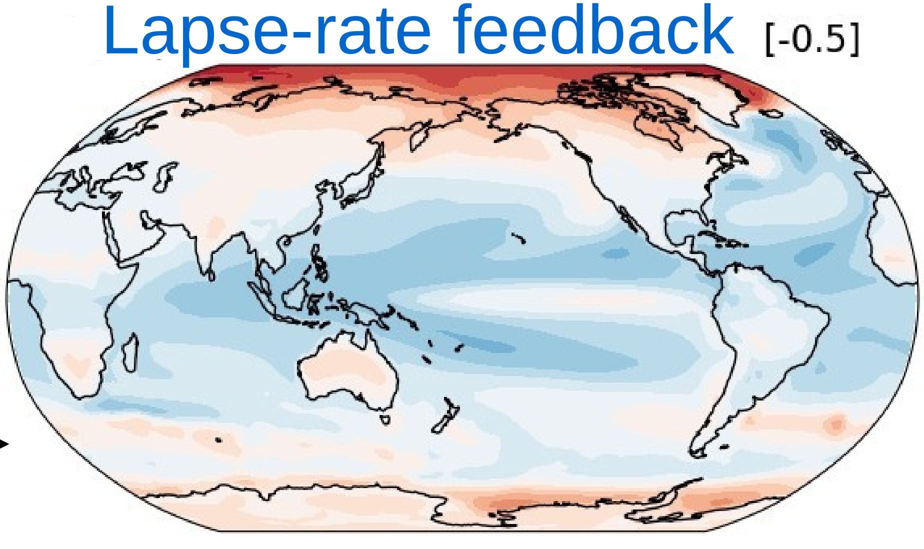


Temperature feedback



Lapse-rate feedback





[courtesy of M. Zelinka 2021]
(https://doi.org/10.5281/zenodo.5206851)

Climate feedbacks

Classical decomposition (***specific humidity***)

$$\lambda = \lambda_T + \lambda_W + \lambda_C + \lambda_\alpha$$

temper ature | water vapor | clouds | surface albedo

$$\lambda_T = \lambda_P + \lambda_L$$

Uniform vertical temperature change ← Planck

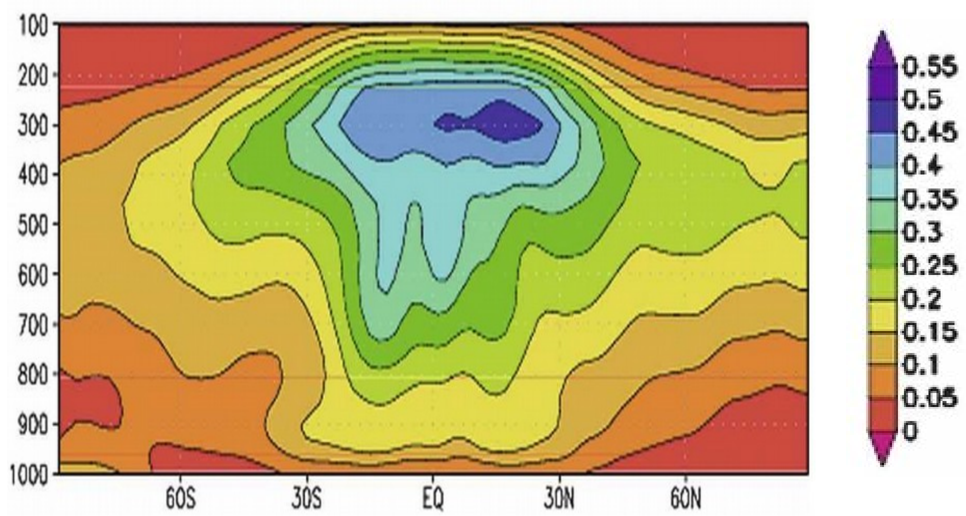
lapse rate → Departure from uniform vertical temperature change

$$\lambda = \frac{dN}{dT_s} = \sum_x \frac{\partial N}{\partial x} \frac{\partial x}{\partial T_s}$$

radiative kernel computed by radiative codes

response to surface temperature change

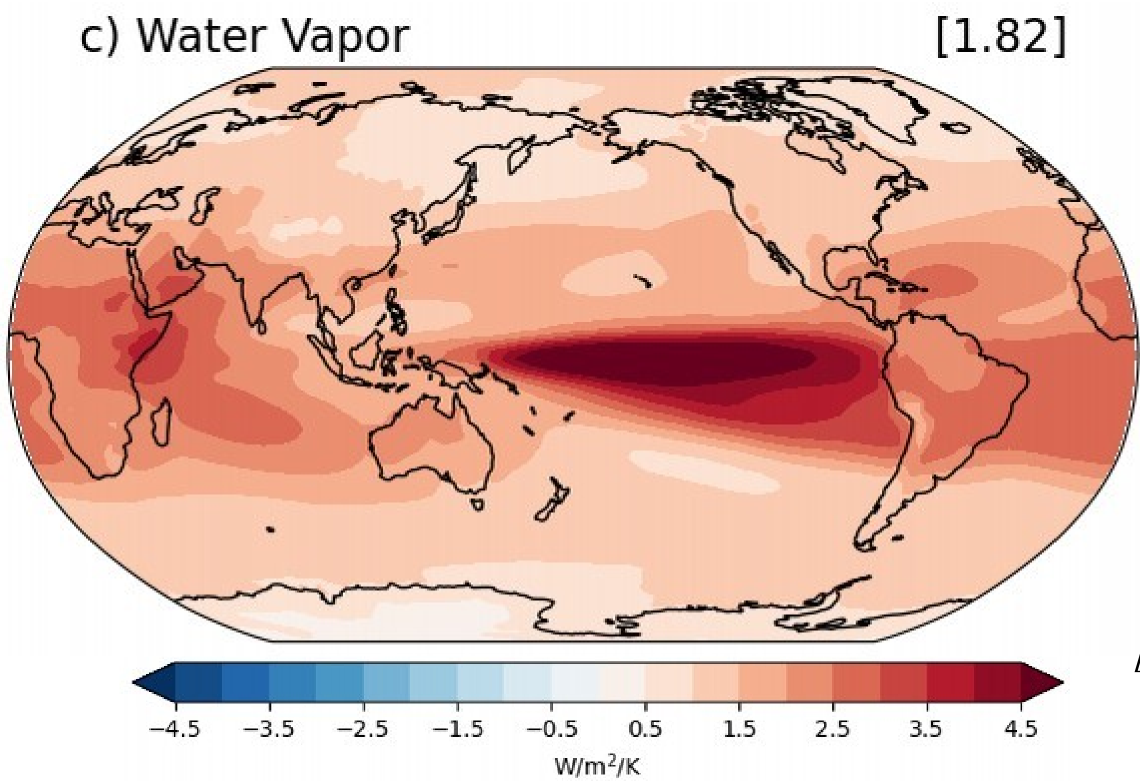
Water vapour feedback



$$\frac{\partial R}{\partial Q_a(P)} \frac{d Q_a(P)}{d T_s}$$

W/m²/K/(100hPa)

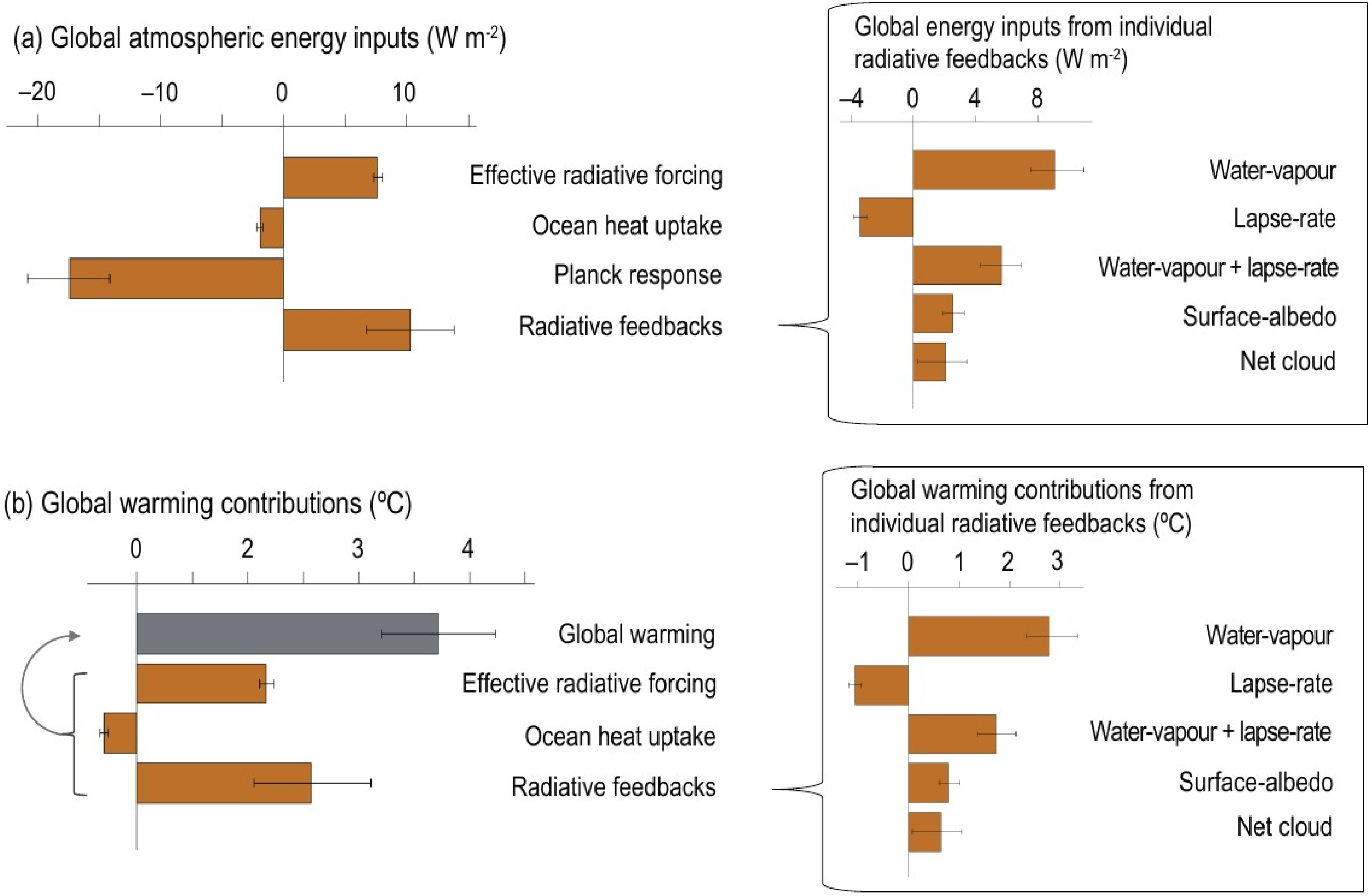
Soden et al., J. Climate, 2008



[courtesy of M. Zelinka 2021]

How much individual feedbacks contribute to global warming

(year 100 after abrupt 4xCO2)



[IPCC AR6 WG1, ch 7]

Emission height and emission temperature

Flux TOA has been emitted by:

the surface — the atmosphere

$$F = \Gamma_s B_s + \int_0^H \frac{\partial \Gamma(z)}{\partial z} B(z) dz$$

where $\mathbf{B}$ is the **Planck** function, $\Gamma_s \equiv \Gamma(0)$ and $\boldsymbol{\Gamma}(\mathbf{z})$ is the hemispherical **transmission function** between altitude z and the TOA:

$$\Gamma(z) = 2 \int_0^1 \exp\left(-\tau(z,\mu)\right) \mu d\mu$$

$\tau(z,\mu)$ is the optical thickness between the TOA and altitude z.

[Dufresne et al. 2020]

Emission height and emission temperature

Flux TOA has been emitted by: the surface, the atmosphere

$$F = \Gamma_s B_s + \int_0^H \frac{\partial \Gamma(z)}{\partial z} B(z) dz$$

May be rewritten as:

$$F = \Gamma_s B_s + (1 - \Gamma_s) B_e$$

$$B_e = \int_0^H B(z) \omega(z) dz$$

$$\omega(z) = \frac{1}{1 - \Gamma_s} \frac{\partial \Gamma(z)}{\partial z}$$

[Dufresne et al. 2020]

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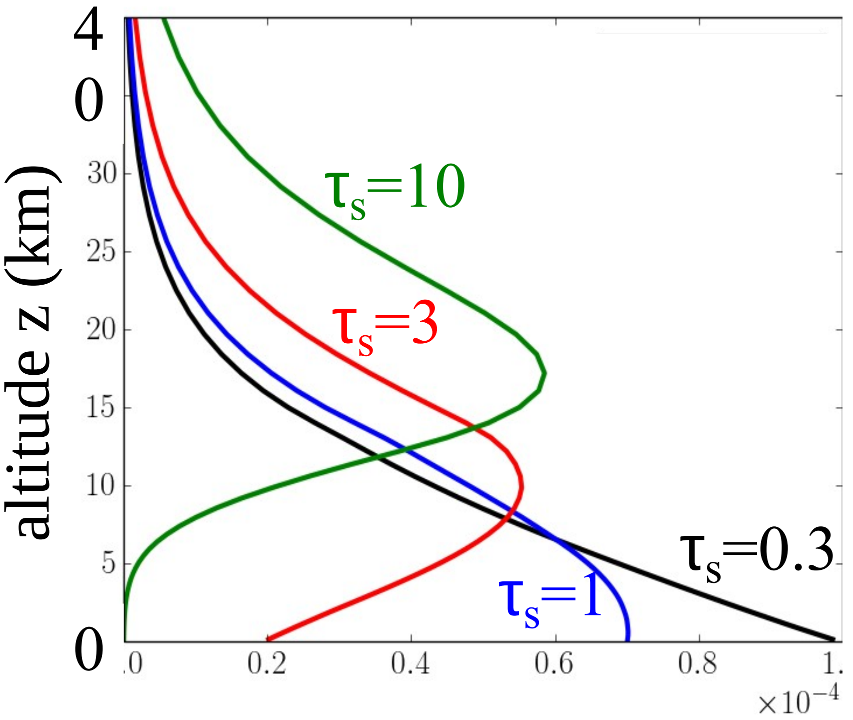
MLS atmospheric profile with a **uniform mass absorption coefficient k** (m2/kg)



ω(z) : optical exchange factor between z and the TOA. Also the ***conditional probability that photons emitted at z reach the TOA***

[Dufresne et al. 2020]

Emission height and emission temperature

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the surface — the atmosphere

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May be rewritten as:

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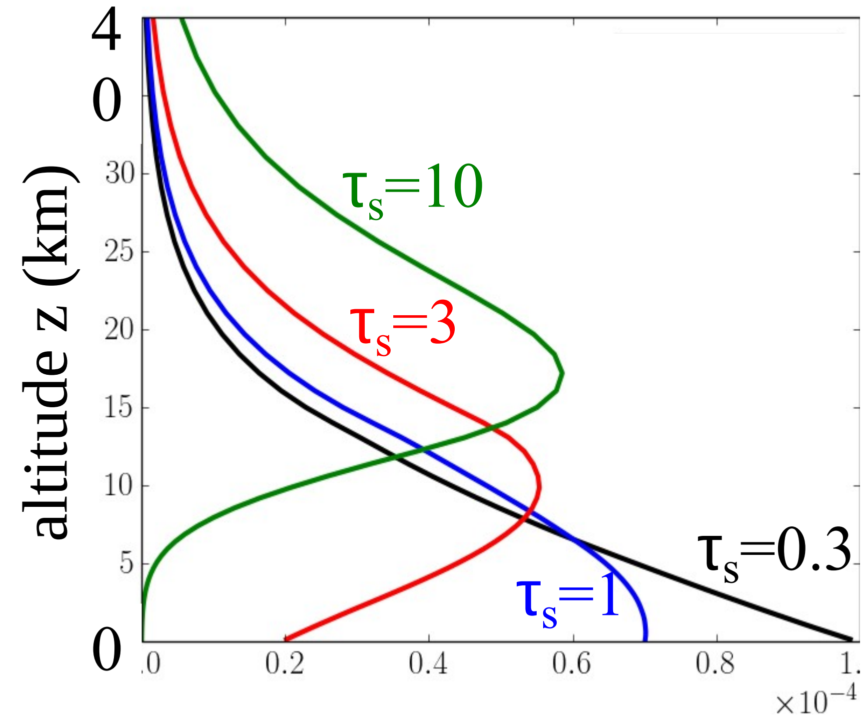
Emission temperature Te: B(Te) = Be

MLS atmospheric profile with a **uniform mass absorption coefficient k** (m2/kg)



A common approximation:
Emission height Ze: altitude where the optical thickness between Ze and the TOA is 1
Emission temperature Te: temperature at altitude Ze

ω(z) : optical exchange factor between z and the TOA. Also the ***conditional probability that photons emitted at z reach the TOA***

[Dufresne et al. 2020]

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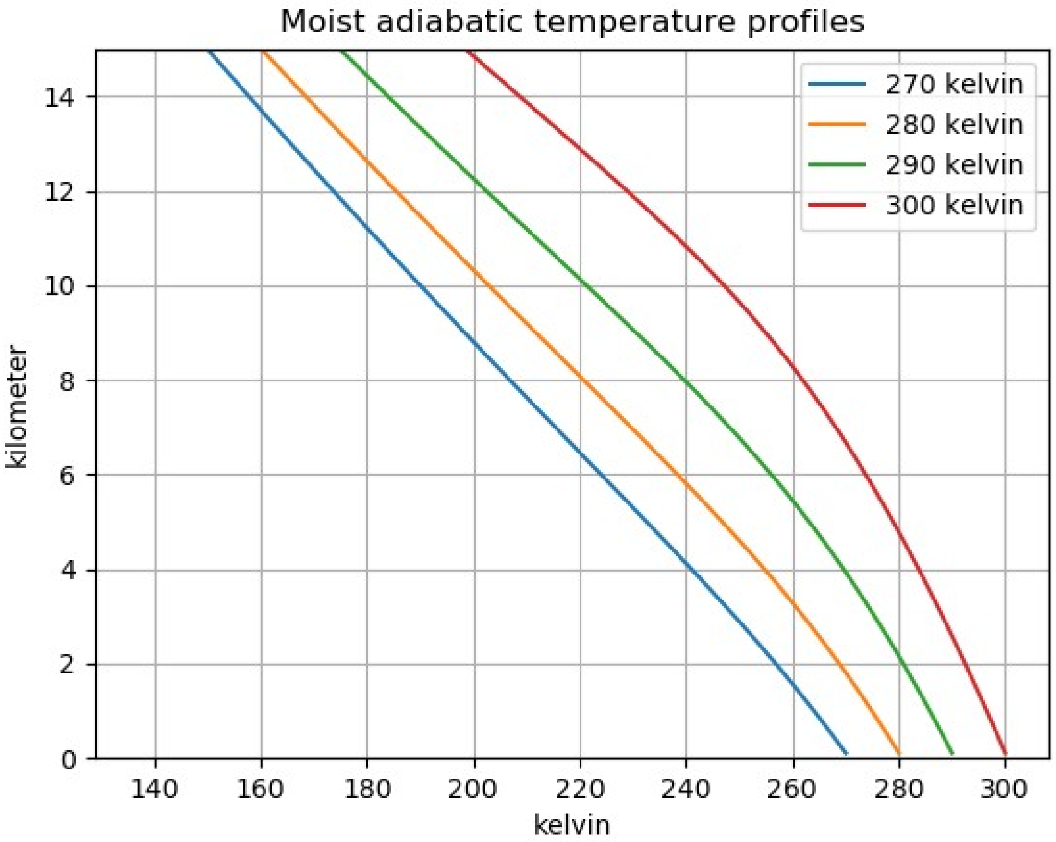
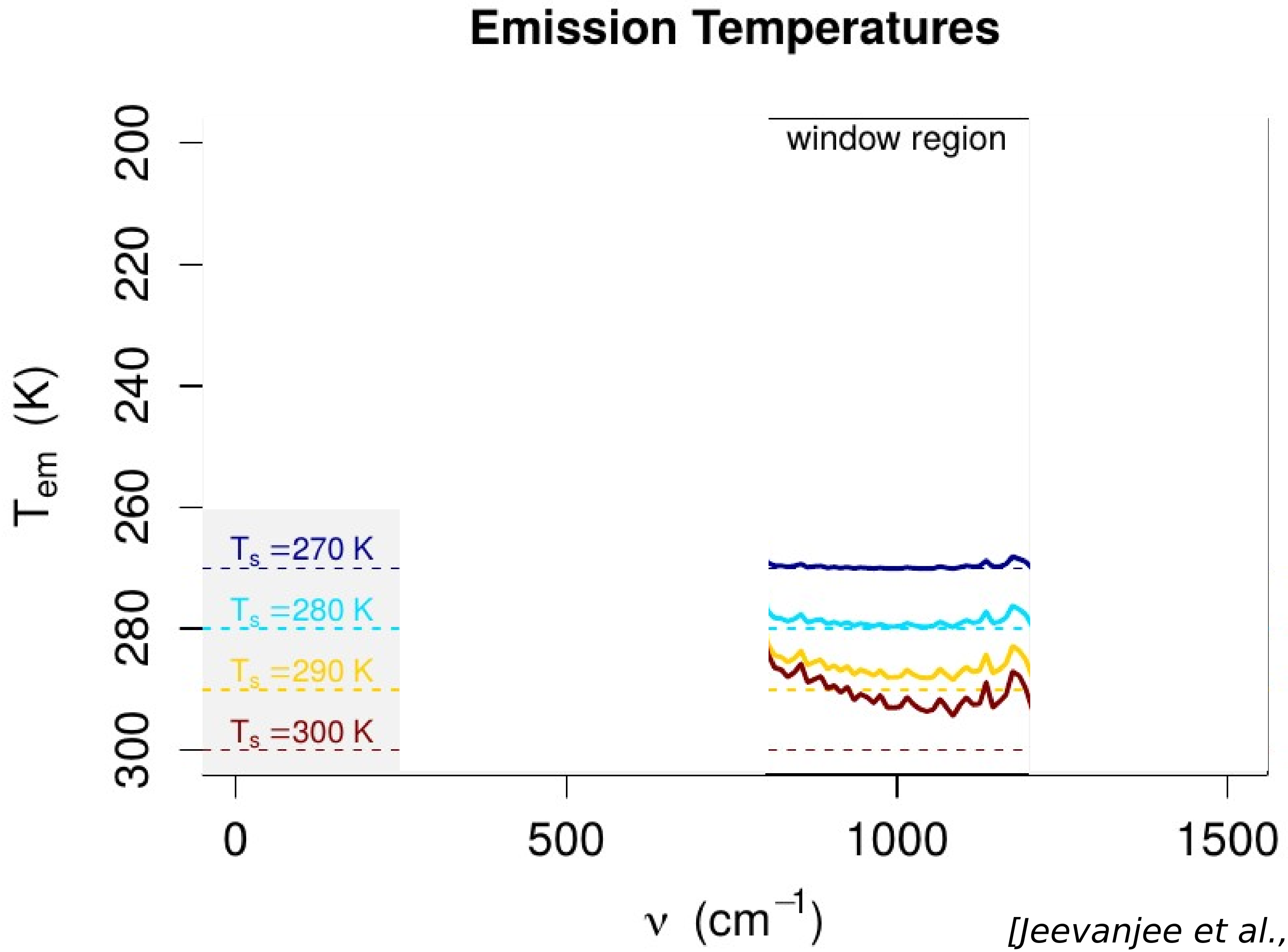
Idealized clear sky tropical atmospheres with ***constant relative humidity***

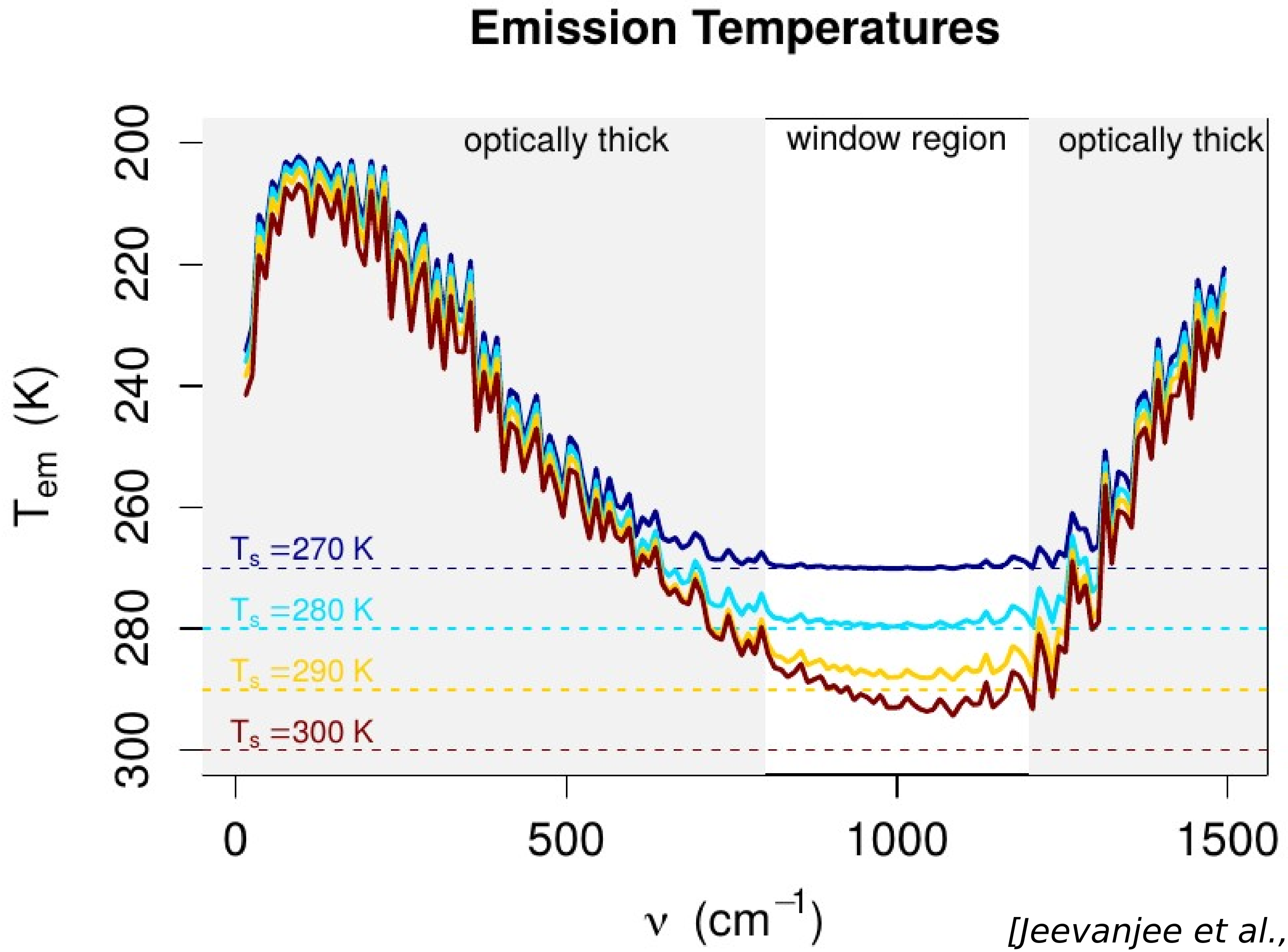


Idealized clear sky tropical atmospheres with ***constant relative humidity***



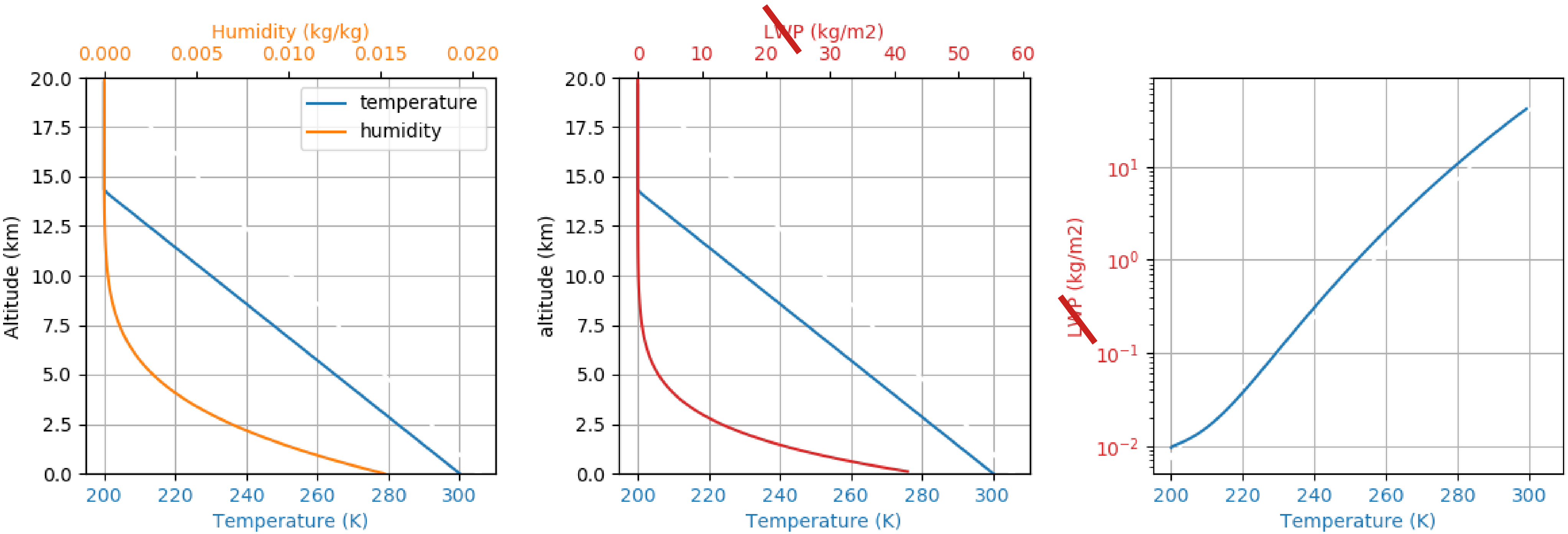
[Jeevanjee et al., GRL, 2021]

*For idealized clear sky tropical atmospheres with **constant relative humidity***



[Jeevanjee et al., GRL, 2021]

Vertical profile of the column integrated amount of water vapour



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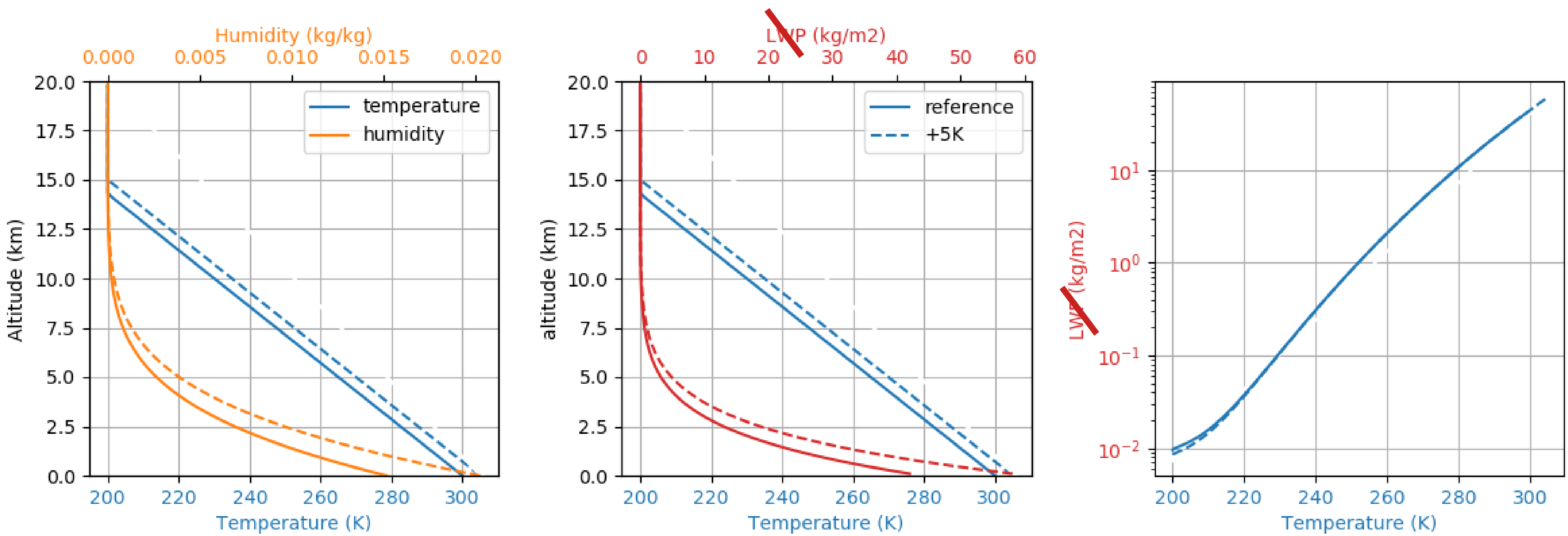
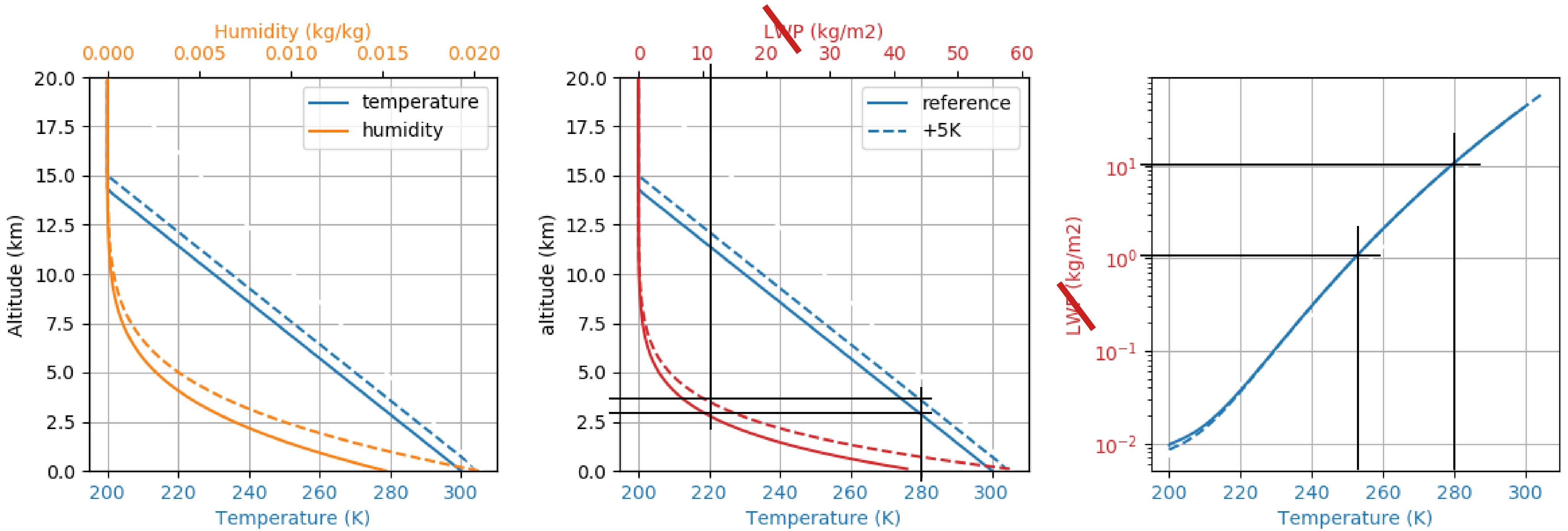


If the relative humidity RH is verticaly uniform in the tropsophere (here 0.7), the **column integrated amount of H_2O** from the TOA to a given altitude is an increasing function than **only depends on the temperature at that altitude**

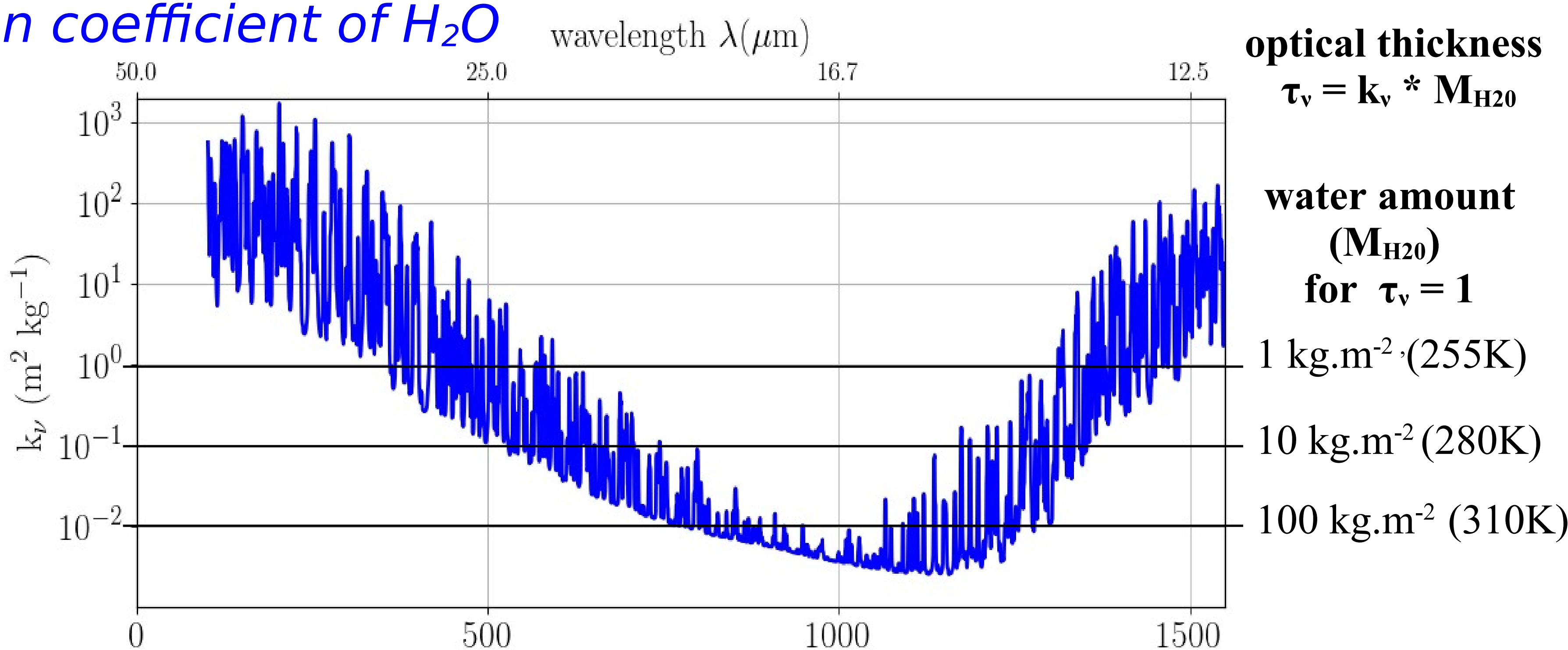
Vertical profile of the column integrated amount of water vapour



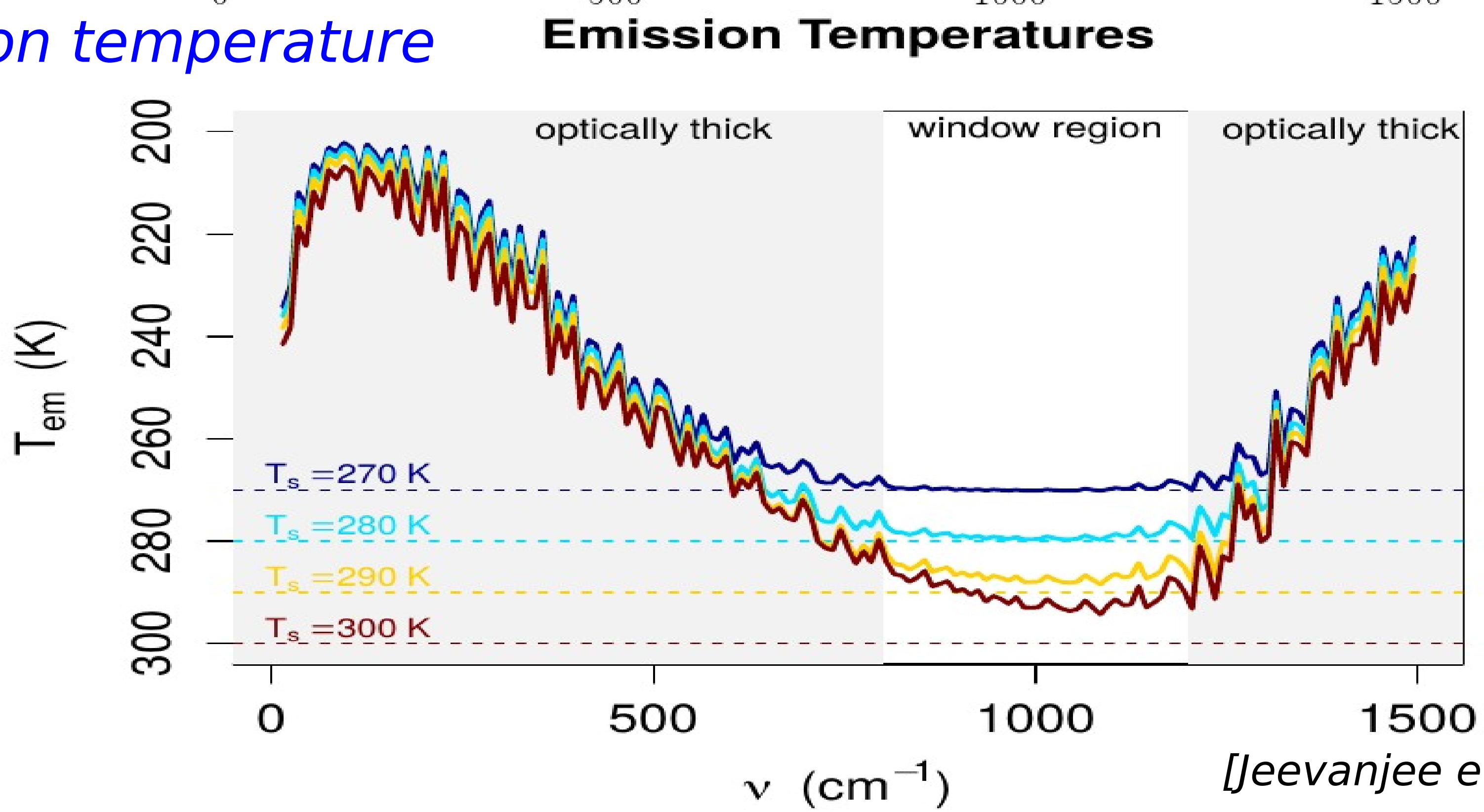
If the relative humidity RH is verticaly uniform in the tropsophere (here 0.7), the column integrated amount of H_2O from the TOA to a given altitude is an increasing function than only depends on the temperature at that altitude

T = 255K => LWP = 1 kg.m^{-2} T = 280K => LWP = 10 kg.m^{-2}

Absorption coefficient of H_2O



Emission temperature

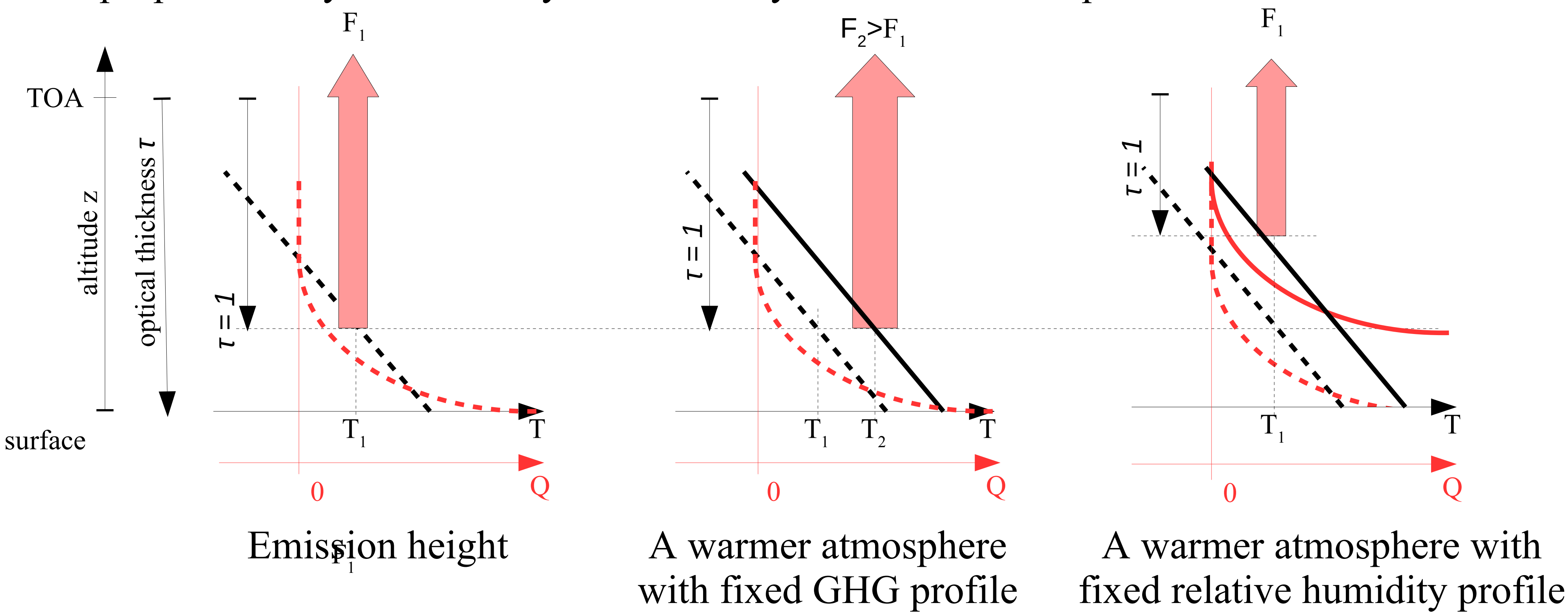


[Jeevanjee et al., GRL, 2021]

Simpson's law

[Simpson 1928, Ingram W. 2010, Jeevanjee et al. 2021]

In spectral regions where gases are highly absorbent of an atmosphere whose properties vary continuously and smoothly with altitude and pressure...



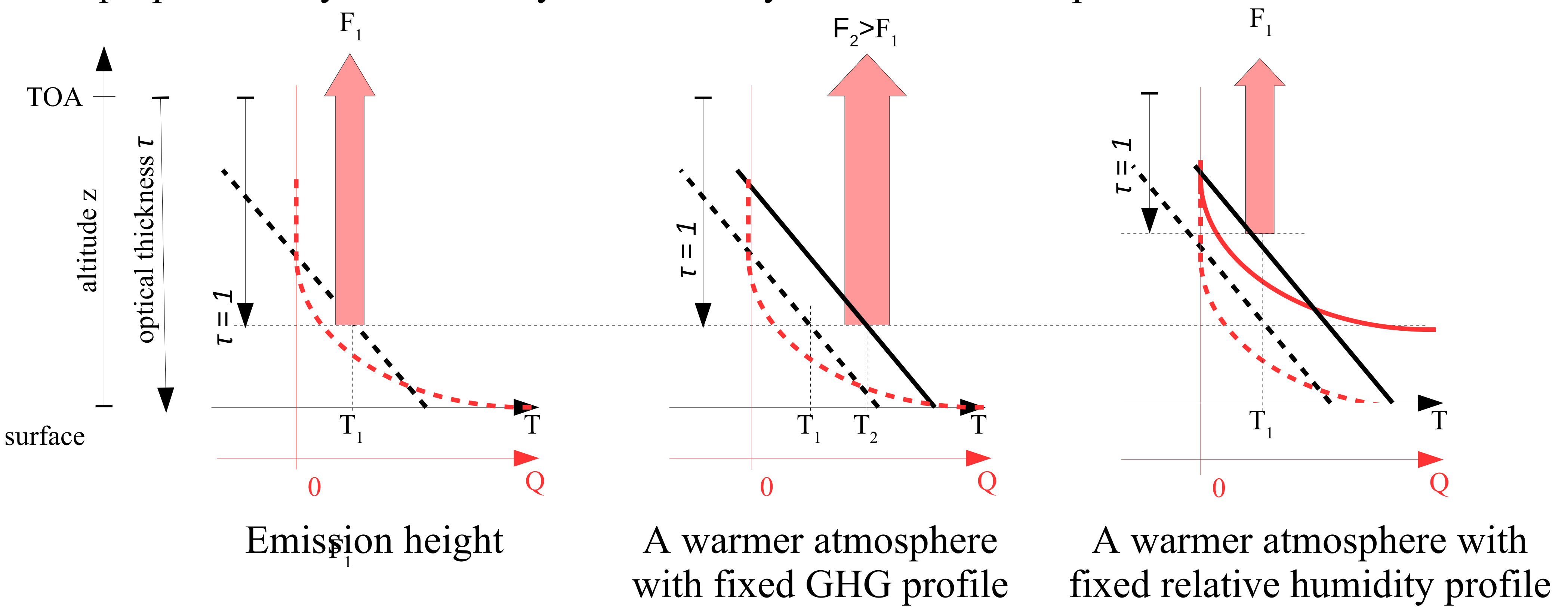
=> the emission temperature (the flux emitted) does not change with a uniform increase in temperature if the relative humidity remains const

=> If these conditions are met over the entire infrared domain, the feedback is infinite, we have a runaway effect

Simpson's law

[Simpson 1928, Ingram W. 2010, Jeevanjee et al. 2021]

In spectral regions where gases are highly absorbent of an atmosphere whose properties vary continuously and smoothly with altitude and pressure...



=> the emission temperature (the flux emitted) does not change with a uniform increase in temperature if the relative humidity remains constant

Key hypothesis:

- Spectral domain with H_2O strong absorption, no other GHGs, no clouds
- Relative humidity is verticaly uniform, as is the temperature change
- The gas absorption properties do not depend on temperature and pressure

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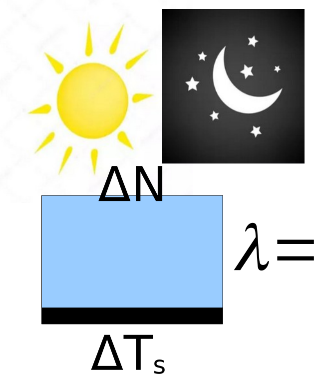
Climate feedbacks



$$\lambda = \frac{dN}{dT_s}$$

Classical decomposition (*specific humidity*)

$$\lambda = \lambda_P + \lambda_L + \lambda_W + \lambda_C + \lambda_\alpha$$

Planck; lapse rate; water vapor; clouds; surface albedo

Relative humidity decomposition (Held & Shell, 2012)

$$\underbrace{\lambda_W}_{\text{water vapor}} = \underbrace{\lambda_{QP} + \lambda_{QL}}_{} + \lambda_R$$

λ_R ← Change in relative himidity

At constante relative humidity, change in humidity du to change, in verically uniform temperature (λ_{QP}) and lapse rate (λ_{QL})

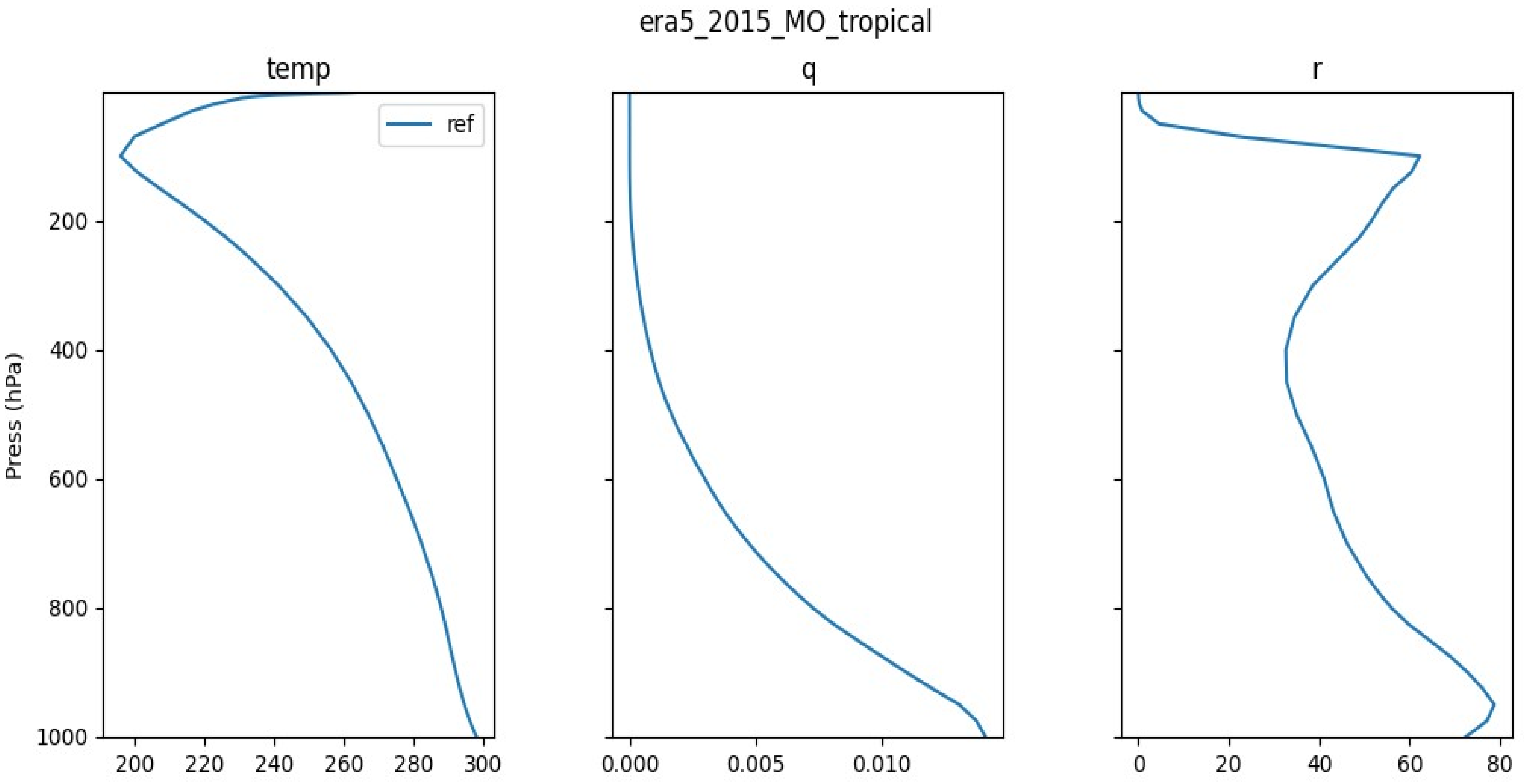
$$\lambda = \underbrace{\lambda_P^* + \lambda_L^*}_{\text{at constante relative humidity}} + \lambda_R + \lambda_C + \lambda_\alpha$$

Planck; lapse rate; relative humidity; clouds; surface albedo

$$\lambda_P^* = \lambda_P + \lambda_{QP}$$

$$\lambda_L^* \doteq \lambda_L + \lambda_{QL}$$

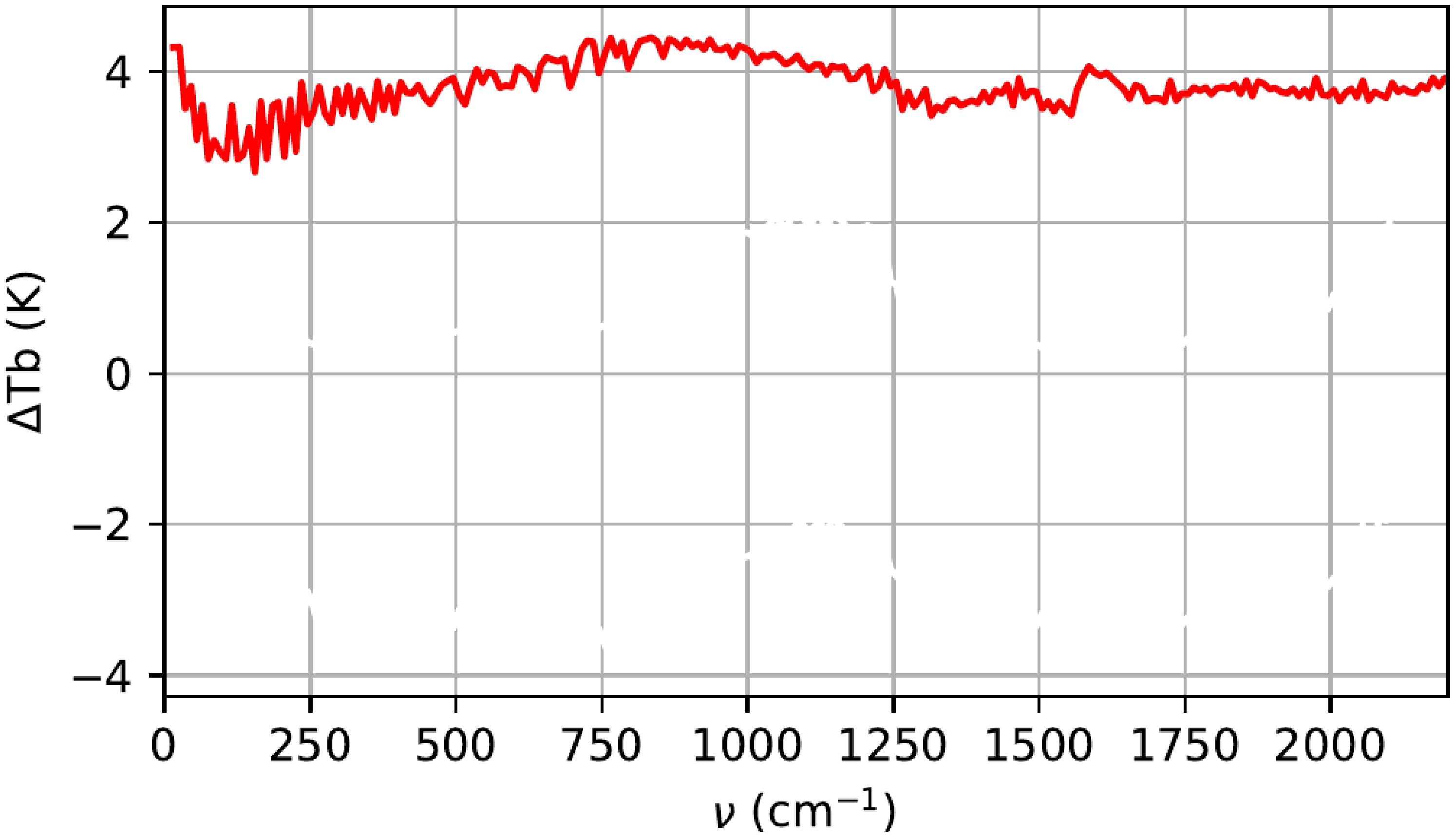
For a mean tropical atmosphere (ERA5)



For a mean tropical atmosphere (ERA5)

Change in brightness temperature
in response to a verticaly **uniform +4K increase of the troposphere**

GHG: Only $H_2 0$



at constant **specific** humidity

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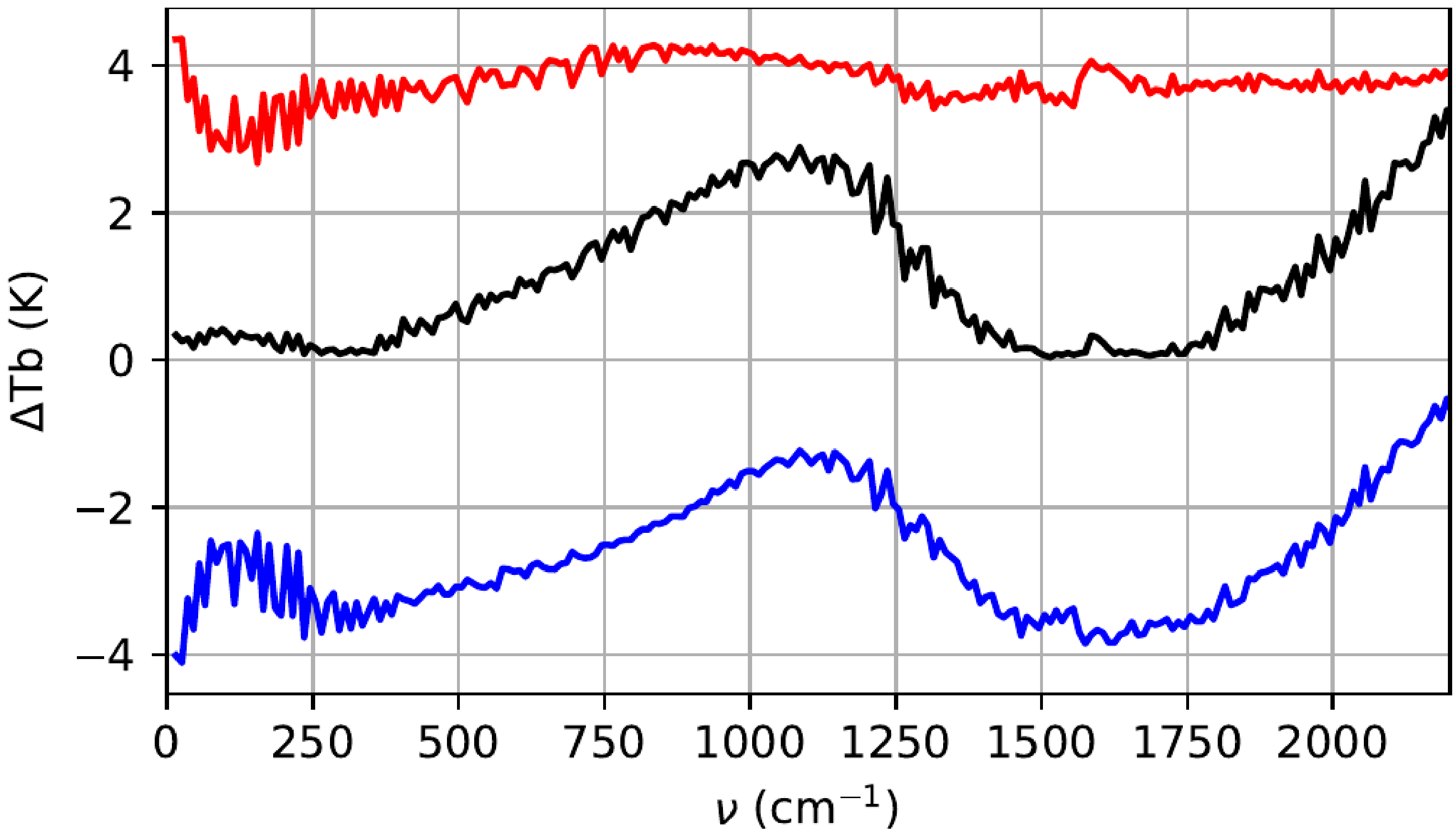


at constant **specific** humidity

at constant **relative** humidity

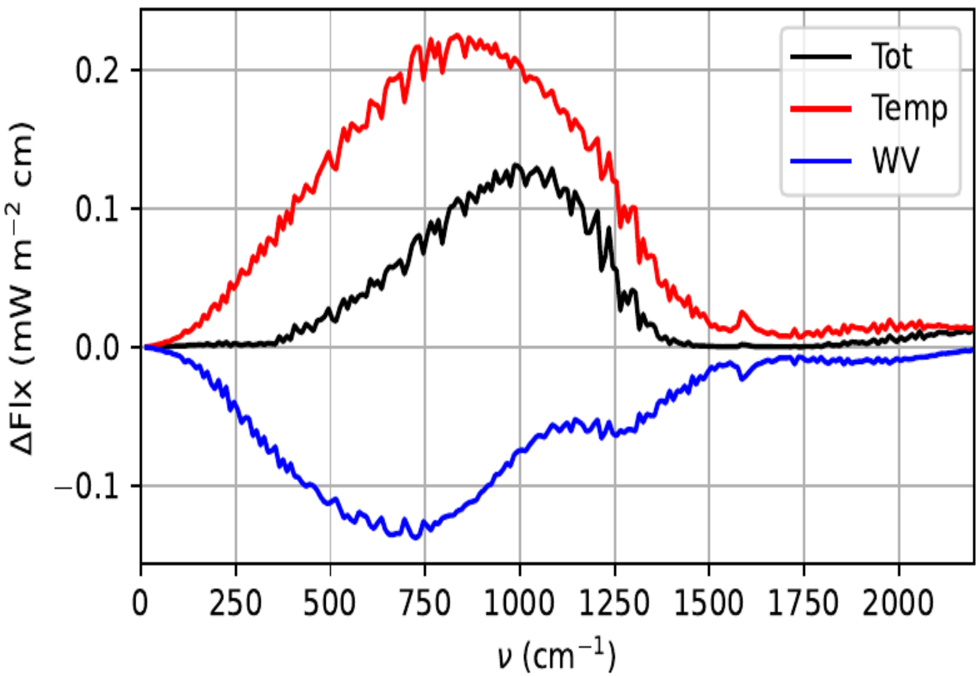
due to change in q **to keep relative humidity constant**

For a mean tropical atmosphere (ERA5)

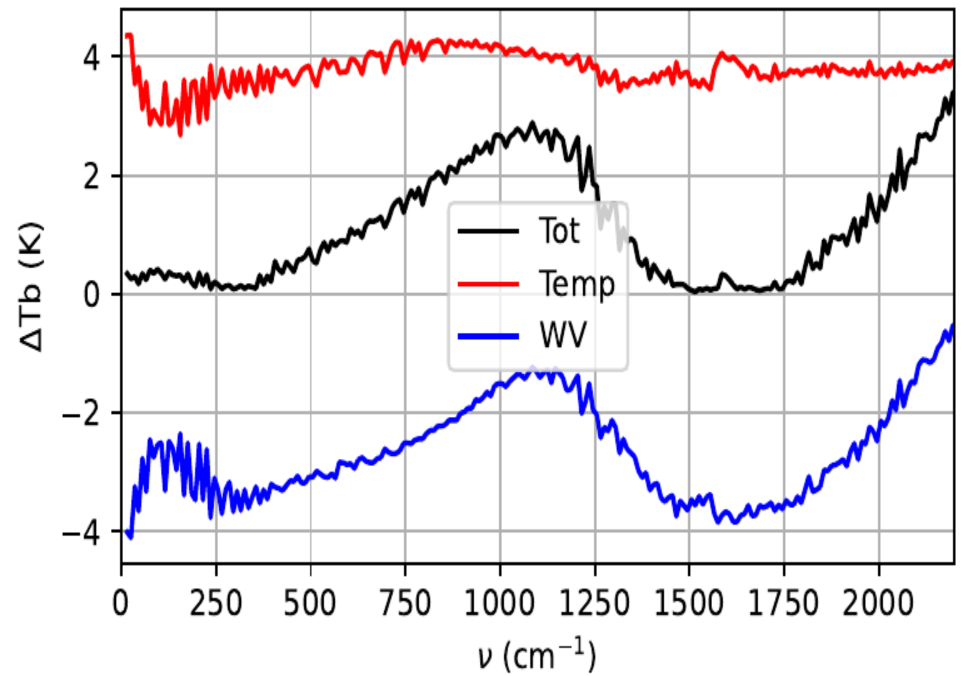
Change in brightness temperature
in response to a verticaly **uniform +4K increase of the troposphere**

GHG: Only H_20

Flux at the TOA



Brightness temperature

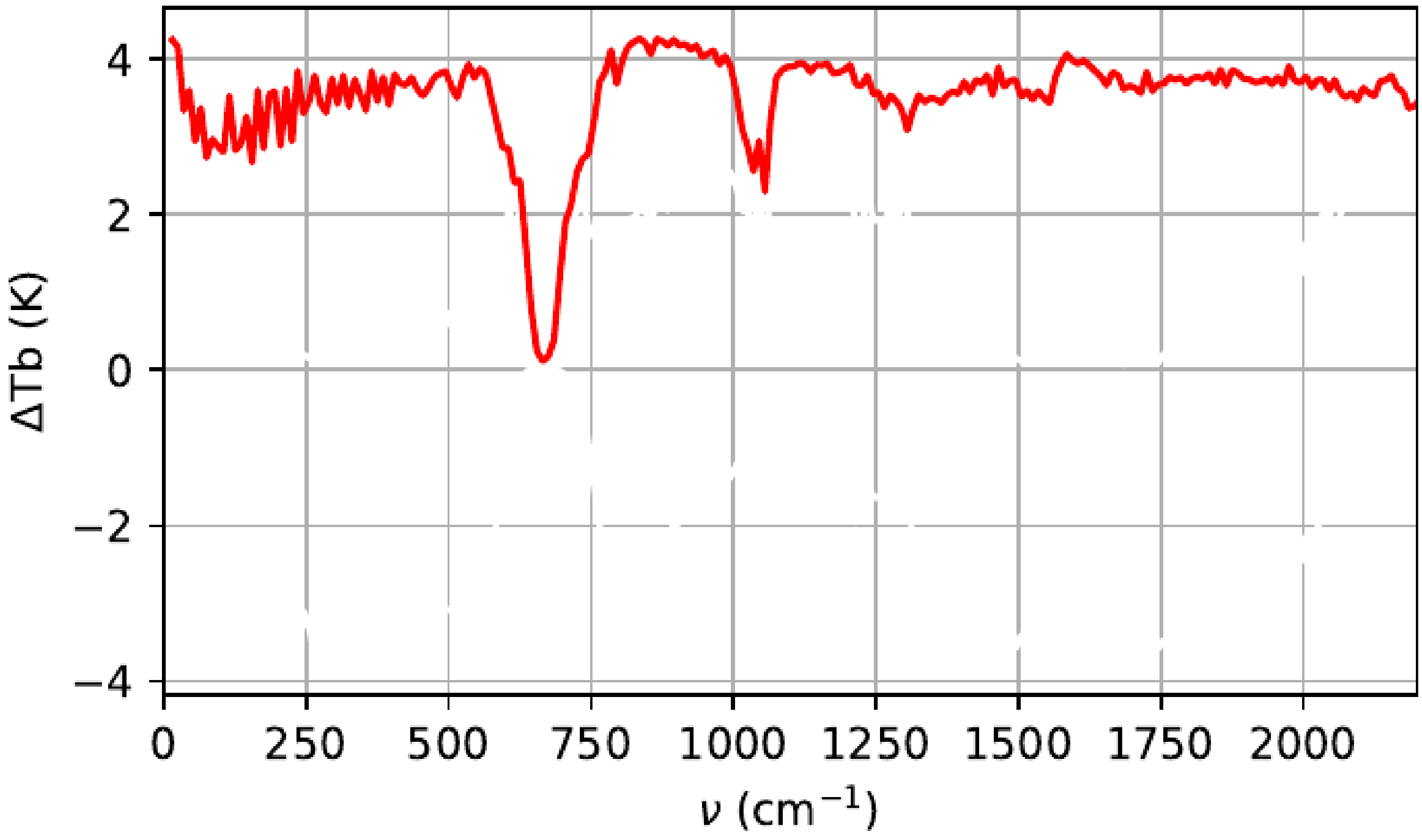


For a mean tropical atmosphere with only H_2O, the **increase in water vapour to keep the relative humidity constant reduces by 60%** the increase in outgoing radiative flux due to the increase in temperature if the specific humidity would remain constant.

For a mean tropical atmosphere (ERA5)

Change in brightness temperature
in response to a verticaly **uniform +4K increase of the troposphere**

GHG: Only $H_20 + CO_2 + O_3$

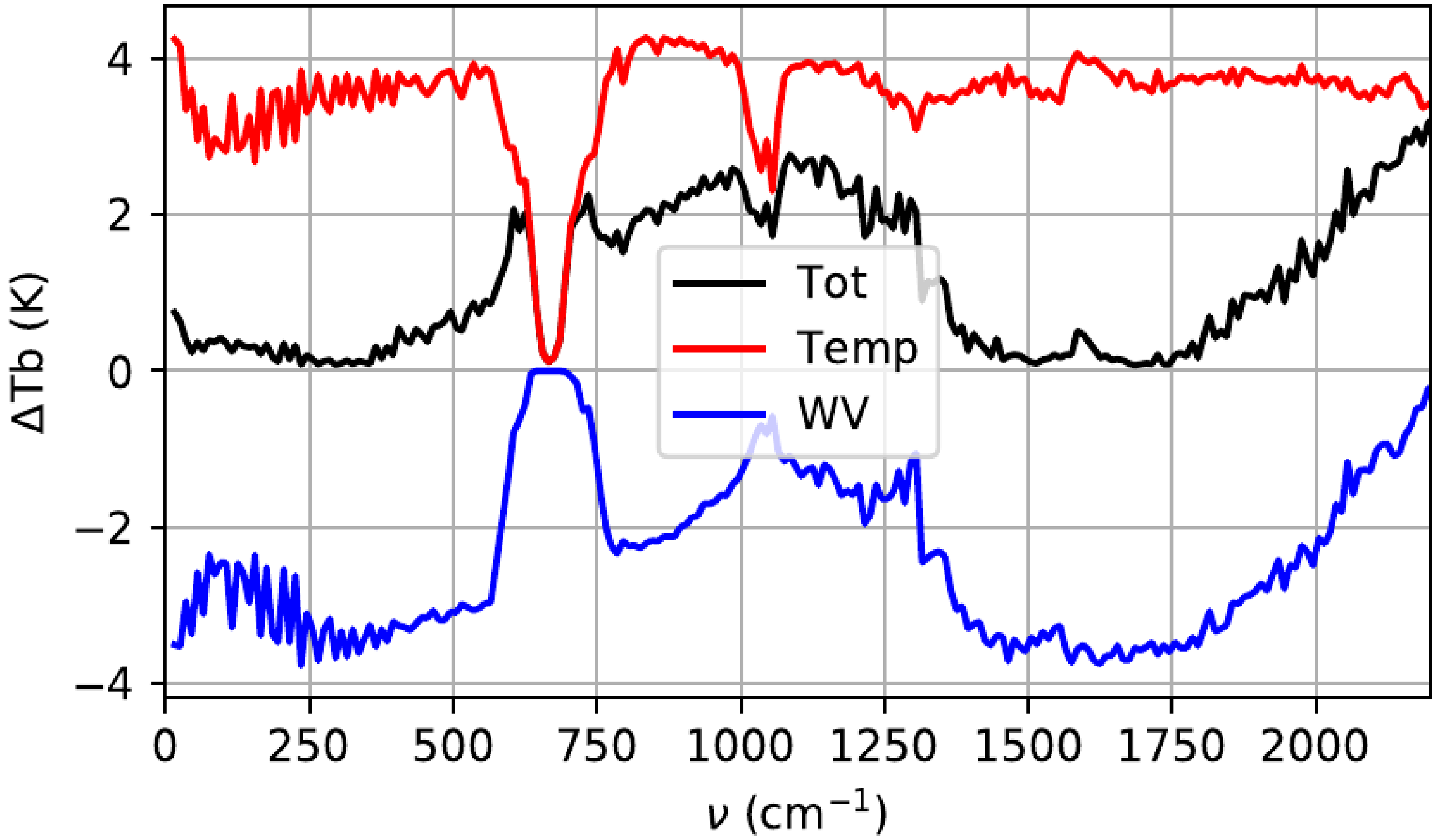


at constant **specific** humidity

For a mean tropical atmosphere (ERA5)

Change in brightness temperature
in response to a verticaly **uniform +4K increase of the troposphere**

GHG: Only $H_20 + CO_2 + O_3$



at constant **specific** humidity

at constant **relative** humidity

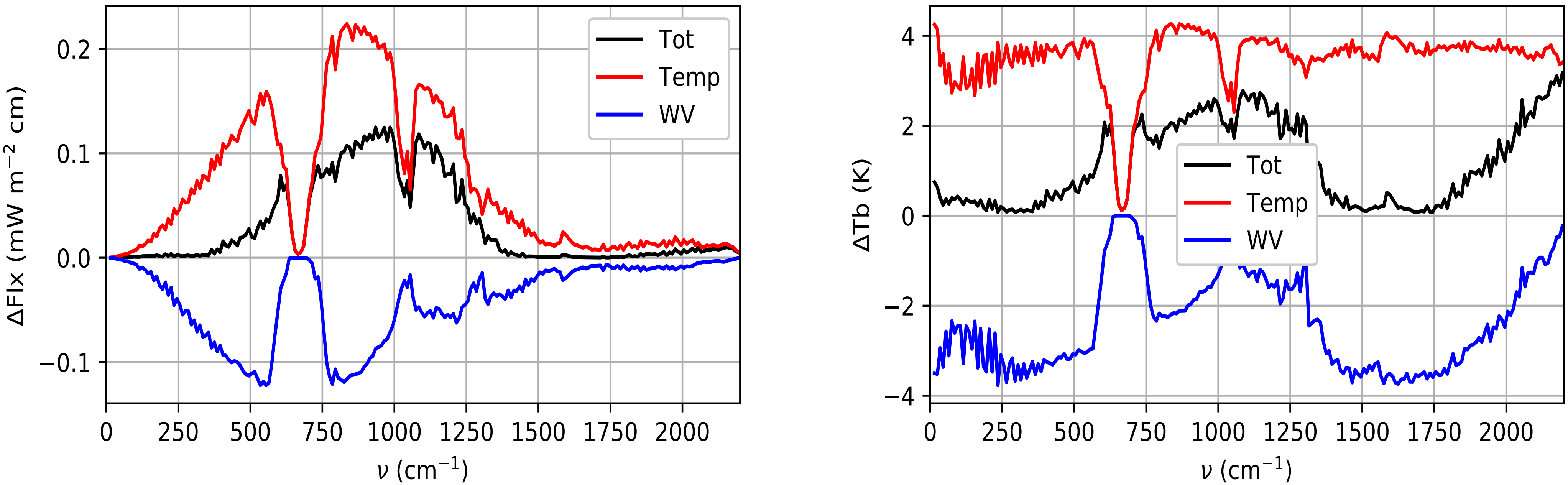
due to change in q **to keep relative humidity constant**

For a mean tropical atmosphere (ERA5)

Change in response to a verticaly **uniform +4K increase of the troposphere**

GHG: Only $H_20 + CO_2 + O_3$

Flux at the TOA　　　　Brightness temperature

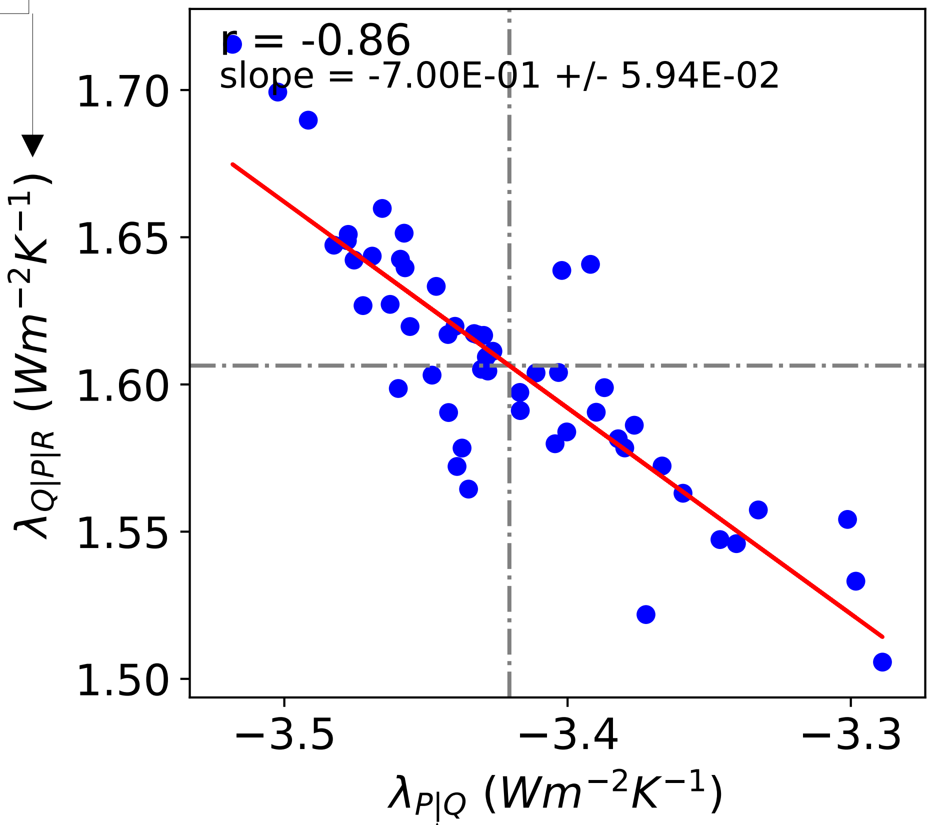


For a mean tropical atmosphere, the **increase in water vapour to keep the relative humidity constant reduces by 56%** the increase in outgoing radiative flux due to the increase in temperature if the specific humidity would remain constant.

Multimodel analysis *(50 CMIP6 models)*

Clear sky, whole globe

Sensitivity of TOA flux to water vapour change to keep the relative humidity constant



- On average, the increase in water vapour to keep the relative humidity constant reduces the Planck response by 47%
- This partial cancelation is also present when looking at the spread among models

Sensitivity of TOA flux to **vertically uniform temperature change at constant specific humidity** (Planck response)

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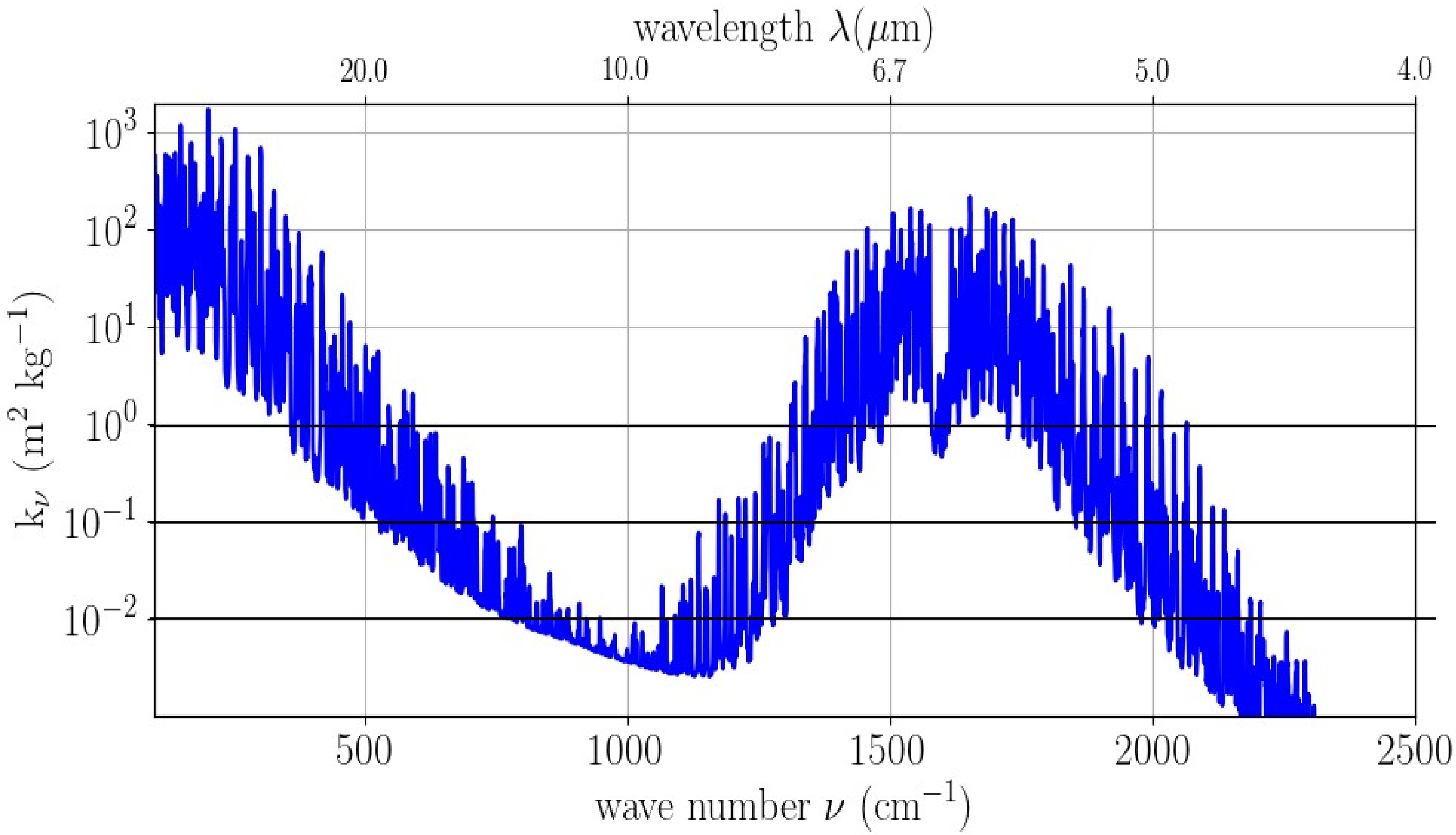
Absorption coefficient of H_2O

optical thickness
Tau = κ * M_{H2O}



water amount for optical thickness tau=1

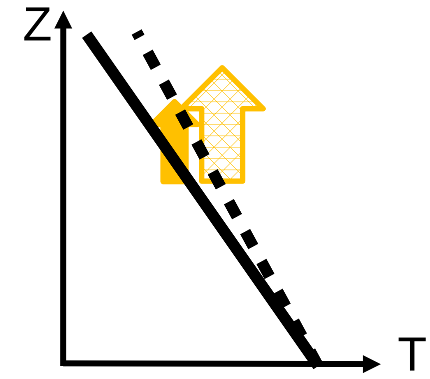
1 kg.m^{-2} (255K)

10 kg.m^{-2} (280K)

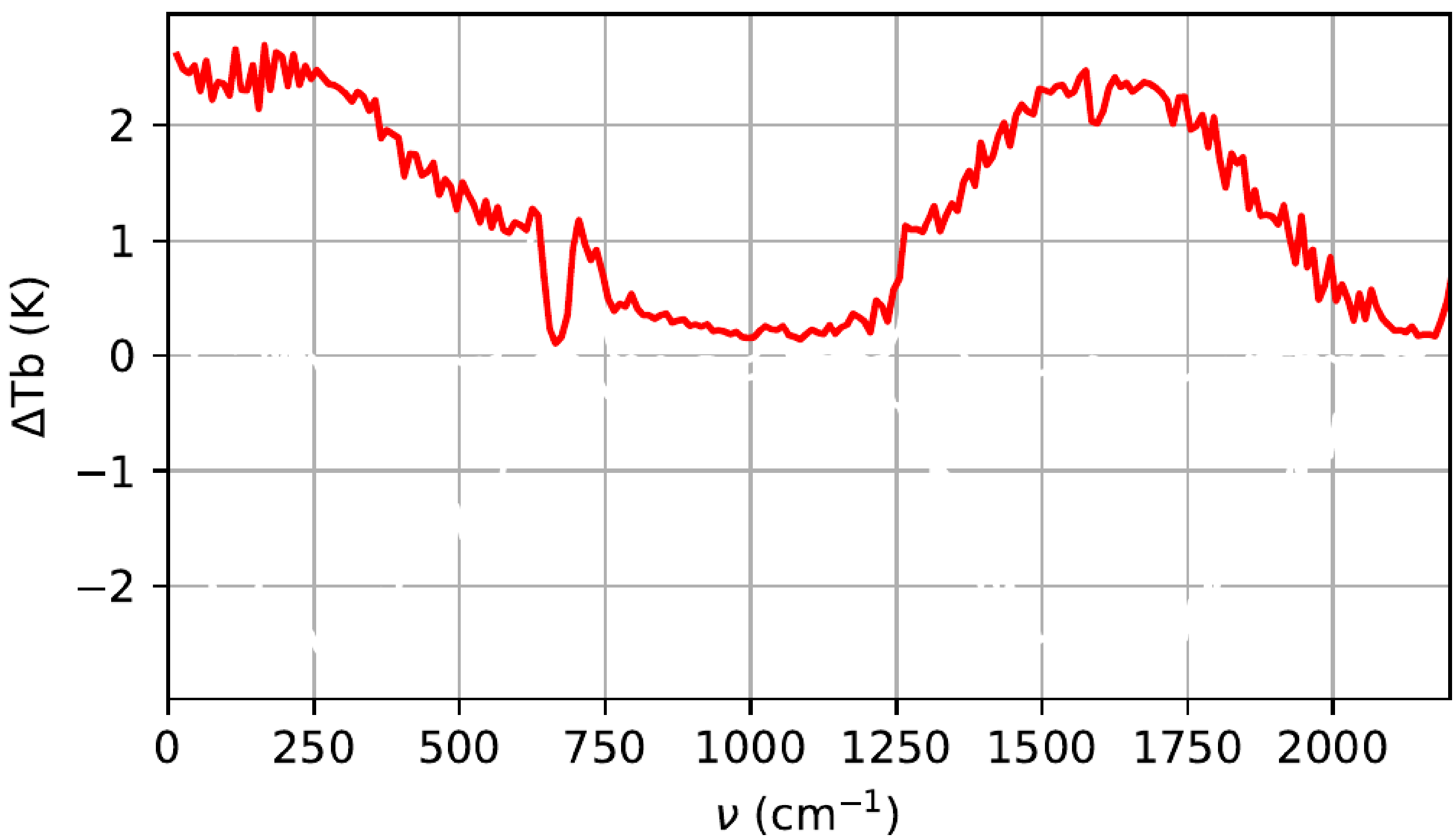
100 kg.m^{-2} (310K)

For a mean tropical atmosphere (ERA5)

Change in brightness temperature
in response to a **change in vertical temperature gradient** (lapse rate):
0 at surface, +4K at the tropopause, $H_2O+CO_2+O_3$



at constant **specific** humidity Q



For a mean tropical atmosphere (ERA5)

Change in brightness temperature
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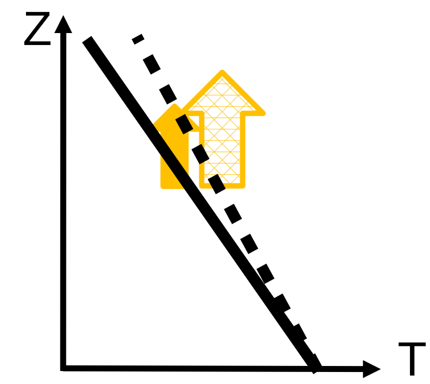
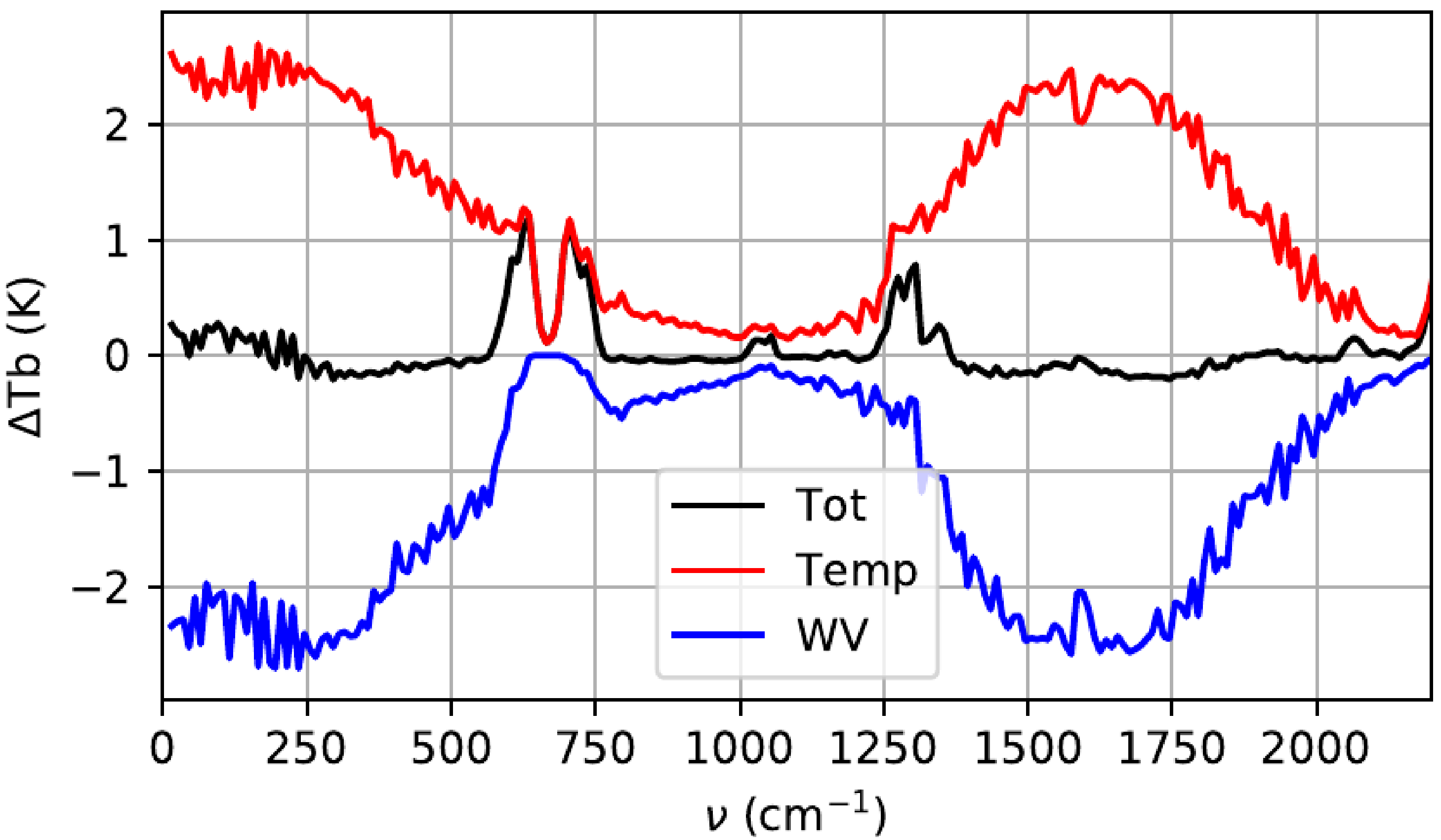
0 at surface, +4K at the tropopause, $H_2O+CO_2+O_3$





at constant **specific** humidity Q

at constant **relative** humidity

due to change in Q **to keep relative humidity constant**

For a mean tropical atmosphere (ERA5)

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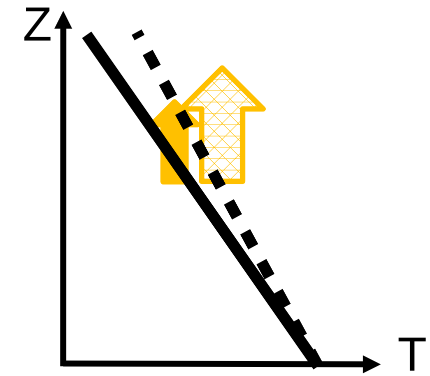
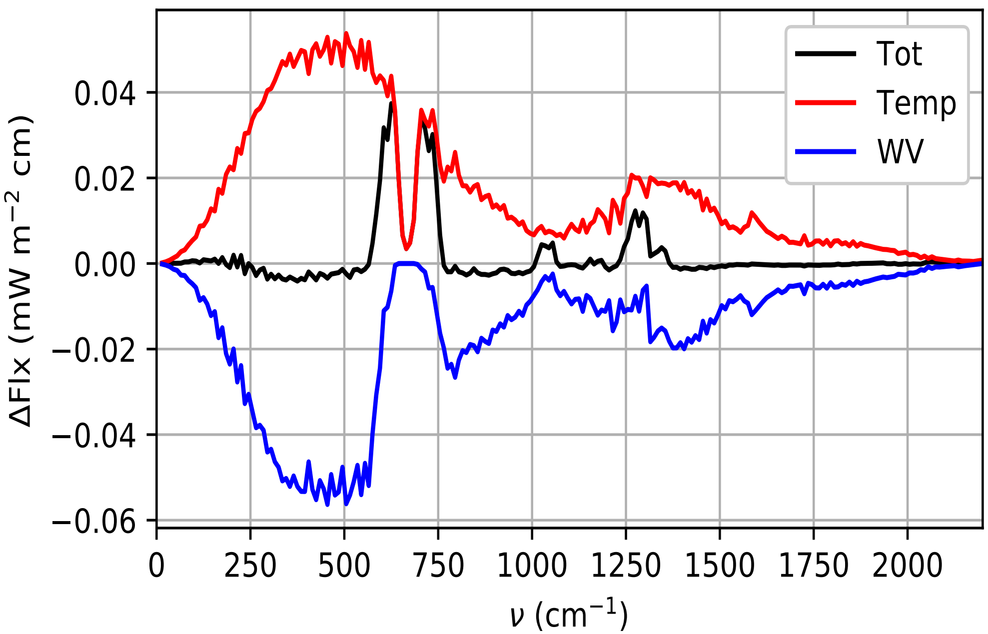
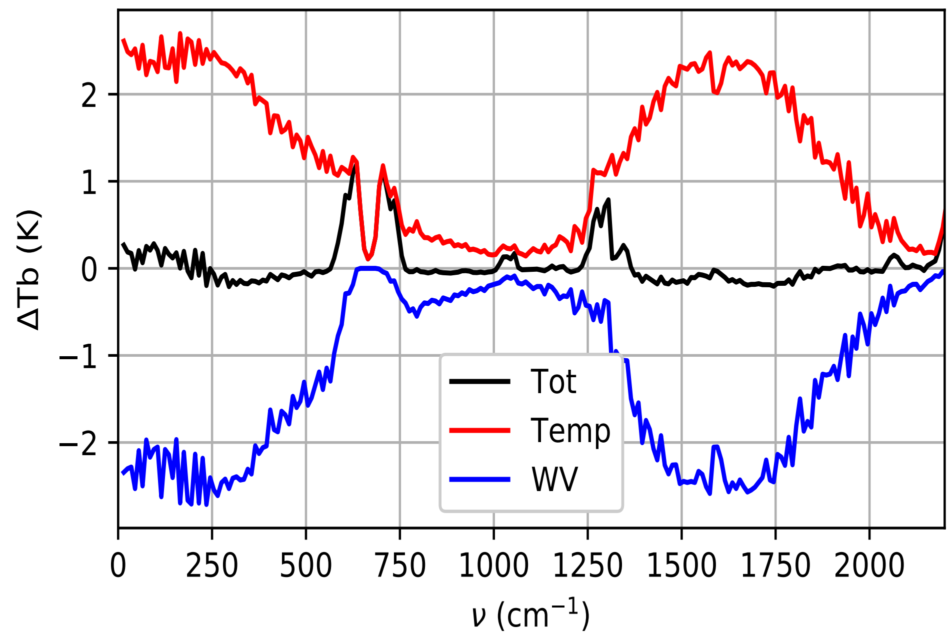


Flux at the TOA



Brightness temperature

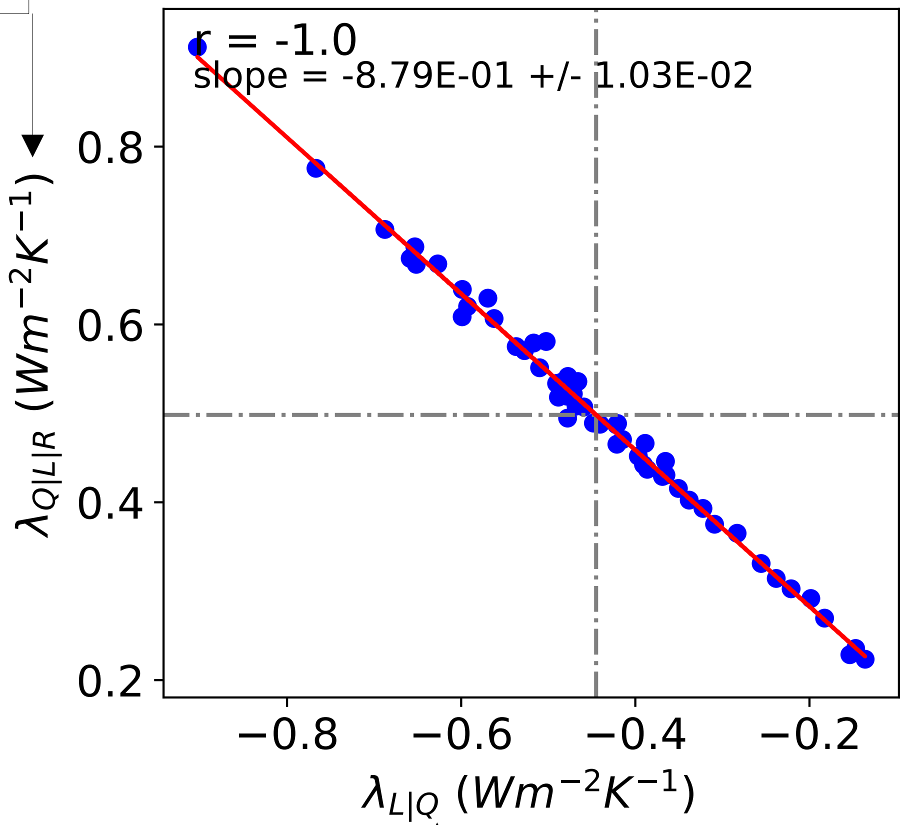


For a mean tropical atmosphere, the **increase in water vapour to keep the relative humidity constant reduces by 90%** the increase in outgoing radiative flux due to change in lapse rate if the specific humidity would remain constant.

Multimodel analysis *(50 CMIP6 models)*

*Clear sky, whole globe, **temperature lapse rate***

Sensitivity of TOA flux to water vapour change to keep the relative humidity constant



- On average, the increase in water vapour to keep the relative humidity constant **more than compensate the lapse rate feedback** (110%)
- A strong cancelation is also present when looking at the spread among models

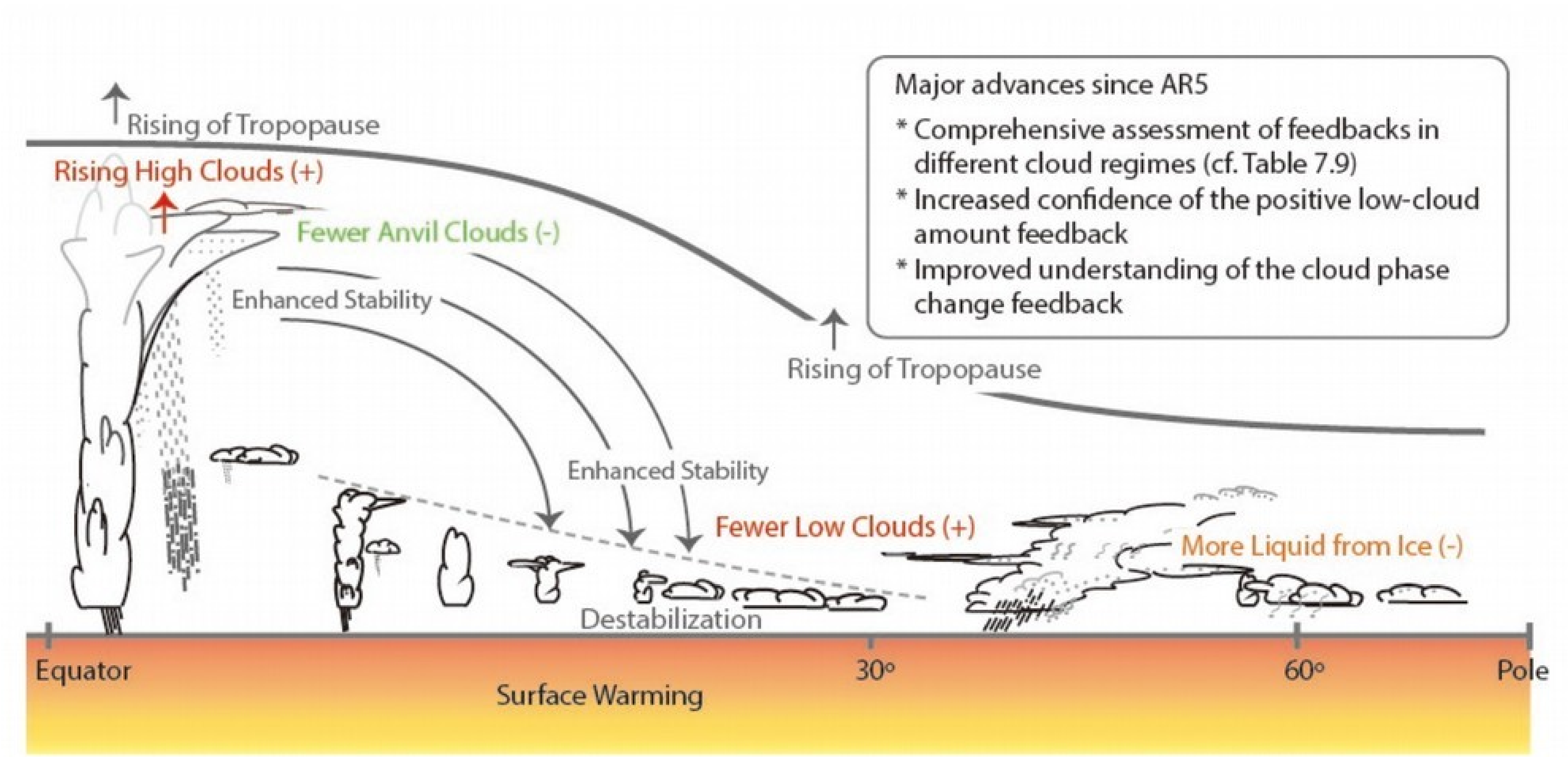
- **The clear sky lapse rate feedback at fixed relative humidity in the LW is almost nul**

Sensitivity of TOA flux to **change in temperature lapse rate at constant specific humidity**

Outlook

- Background
- Water vapour feedbacks in idealized cases :The “Simpson law”
- Clear-sky water vapour feedbacks in realistic cases
- Implication for the clear-sky lapse-rate feedback
- All sky multi-model results
- Conclusion

Cloud feedbacks



Schematic cross section of diverse cloud responses to surface warming. Thick solid and dashed curves indicate the tropopause and the subtropical inversion layer. Thin grey text and arrows represent robust responses. Text and arrows in red, orange and green show the major cloud responses assessed with high, medium and low confidence, respectively, and the sign of their feedbacks to the surface warming is indicated in the parenthesis.

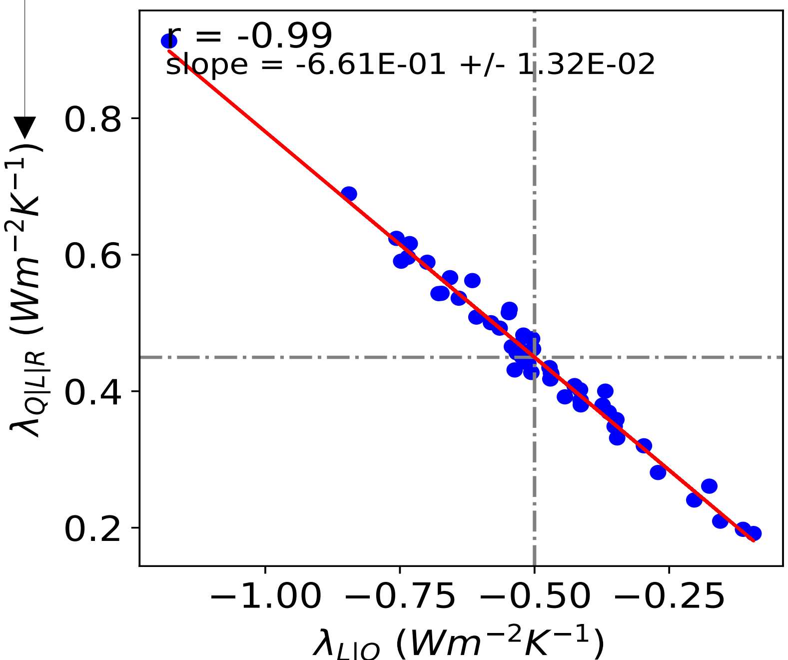
[IPCC AR6 WG1, ch 7]

Multimodel analysis *(50 CMIP6 models)*

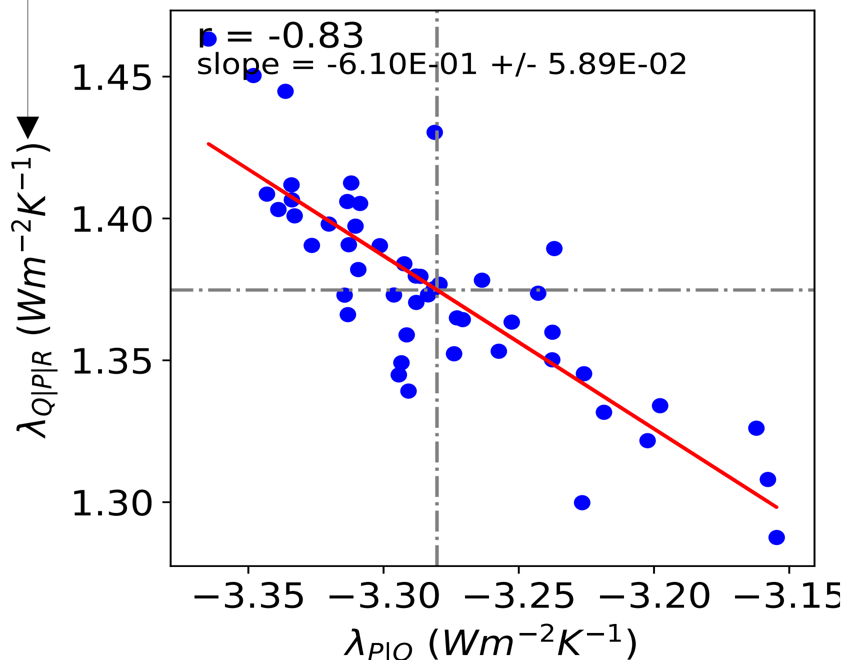
All sky, whole globe

Sensitivity of TOA flux to water vapour change to keep the relative humidity constant

Change in **temperature lapse rate**



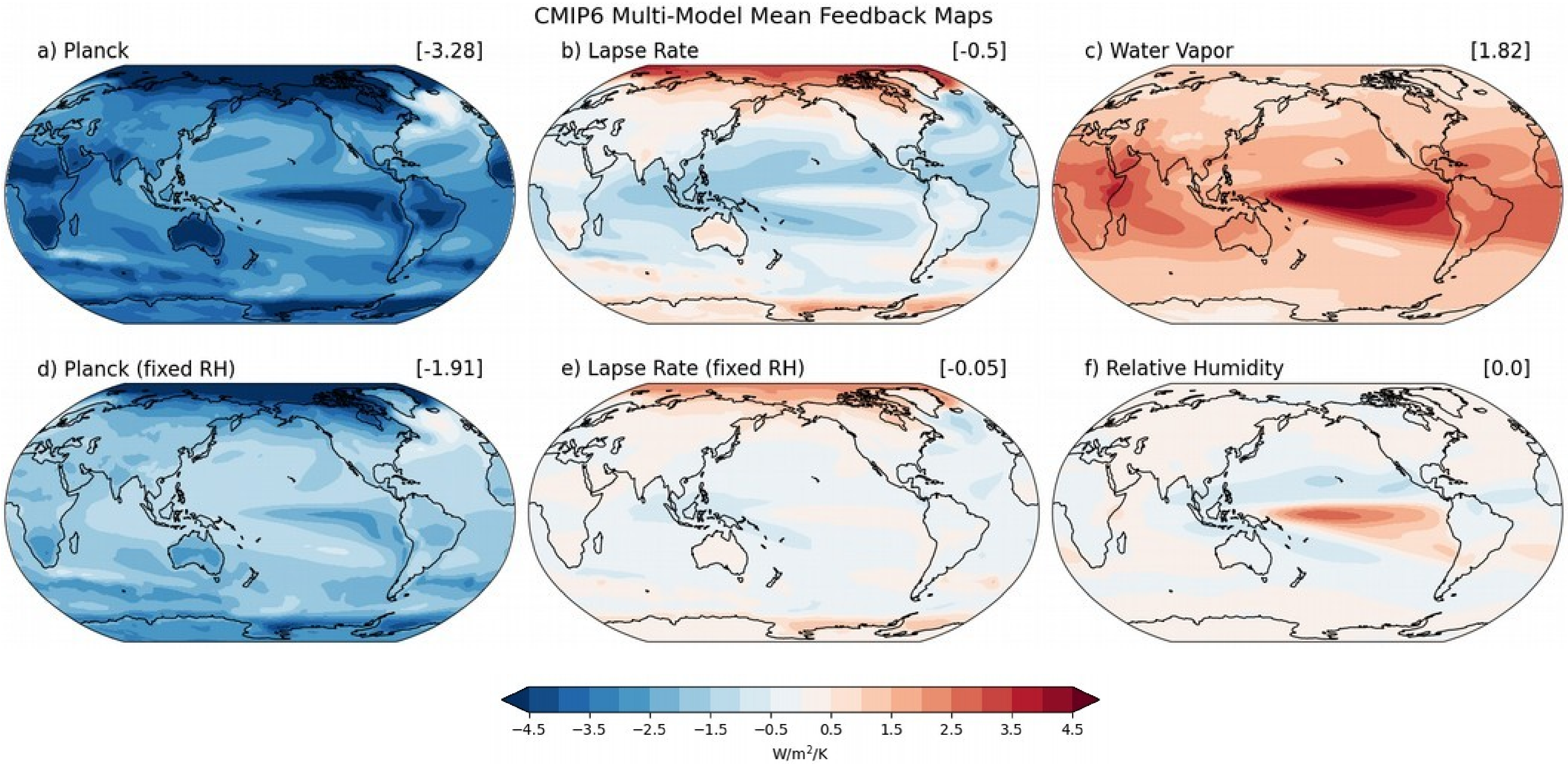
Vertically uniform temp. change (**Planck**)



Sensitivity of TOA flux to **temperature change at constant specific humidity**

- On average, the increase in water vapour to keep the relative humidity constant **reduces the Planck response by 40%**, and the **lapse-rate feedback by 90%**
- This partial cancelation is also present when looking at the spread among models

Climate feedbacks with the absolute and relative humidity decompositions



[courtesy of M. Zelinka 2021]
(https://doi.org/10.5281/zenodo.5206851)

Multimodel analysis *(50 CMIP6 models)*

Climate feedbacks

		absolute humidity decomposition			
	$\lambda_{\theta W}$	$\lambda_{P\|Q}$	$\lambda_{L\|Q}$	$\lambda_{Q\|\theta}$	QSum
mean	-1.95	-3.28	-0.50	1.83	-
std. dev. among models	0.10	0.05	0.19	0.16	0.25

Temperature + WV feedback

Planck response

Lapse-rate feedback

Specific humidity feedback

Square root of quadratic sum

Multimodel analysis *(50 CMIP6 models)*

Climate feedbacks

		absolute humidity decomposition				relative humidity decomposition			
	$\lambda_{\theta W}$	$\lambda_{P\|Q}$	$\lambda_{L\|Q}$	$\lambda_{Q\|\theta}$	QSum	$\lambda_{P\|R}$	$\lambda_{L\|R}$	$\lambda_{R\|\theta}$	QSum
mean	-1.95	-3.28	-0.50	1.83	-	-1.91	-0.05	0.01	-
std. dev. among models	0.10	0.05	0.19	0.16	0.25	0.03	0.07	0.06	0.10
	Temperature + WV feedback	**Planck** response	**Lapse-rate** feedback	**Specific humidity** feedback	Square root of quadratic sum	**Planck** response	**Lapse-rate** feedback	**Relative humidity** feedback	Square root of quadratic sum

Conclusion

- In spectral ranges where H_20 is the main absorber and the atmosphere is optically thick, a change in atmospheric temperature does not induce a change in LW fluxes at the top of the atmosphere if the relative humidity remains constant

- This result is based on the fundamental laws of radiation

- The lapse rate feedback at fixed relative humidity in the LW is almost nul (clear-sky and all sky)

- This allows a simple physically based estimate of the water vapour feedback

- The relative humidity feedback decomposition is much more relevant than the specific humidity decomposition

- Interest of using temperature as “vertical coordinate” instead of pressure (altitude)

This presentation is mainly based on:

- Simpson, G. (1928). Some studies in terrestrial radiation. Memoirs of the Royal Meteorological Society, 2(16), 69–95.
- Jeevanjee, N., Koll, D., & Lutsko, N. (2021). Simpson's law and the spectral cancellation of climate feedbacks. Geophys. Res. Lett..
- Ingram, W. (2010). A very simple model for the water vapour feedback on climate change. QJRMS, 136 (646), 30-40. doi: 10.1002/qj.546
- Held, I. M., & Shell, K. M. (2012, April). Using relative humidity as a state variable in climate feedback analysis. J. Clim., 25 (8), 2578–2582. doi: 10.1175/JCLI-D-11-00721.1
- Ongoing work with M. Zlinka, S. Bony, R. Armante, etc.

Thank you for your attention


Thank you for your attention

- Dans les domaines spectraux où H20 est le principal absorbant et l'atmopshère est optiquement épaisse, un changement de la température de l'atmosphère n'induit pas un changement des flux au sommet de l'atmosphère si l'humidité relative reste constante

- L'utilisation du couple température – humidité relative (T-RH) est plus pertinente celle température – humidité spécidique (T-q)

Analyse multimodèle:

- *Analyse des différences: recherche de problèmes, des origines des désaccords*
- *Analyse des ressemblances: plusieurs planètes légèrement différentes mais obéissant aux mêmes lois physiques*

Conclusion

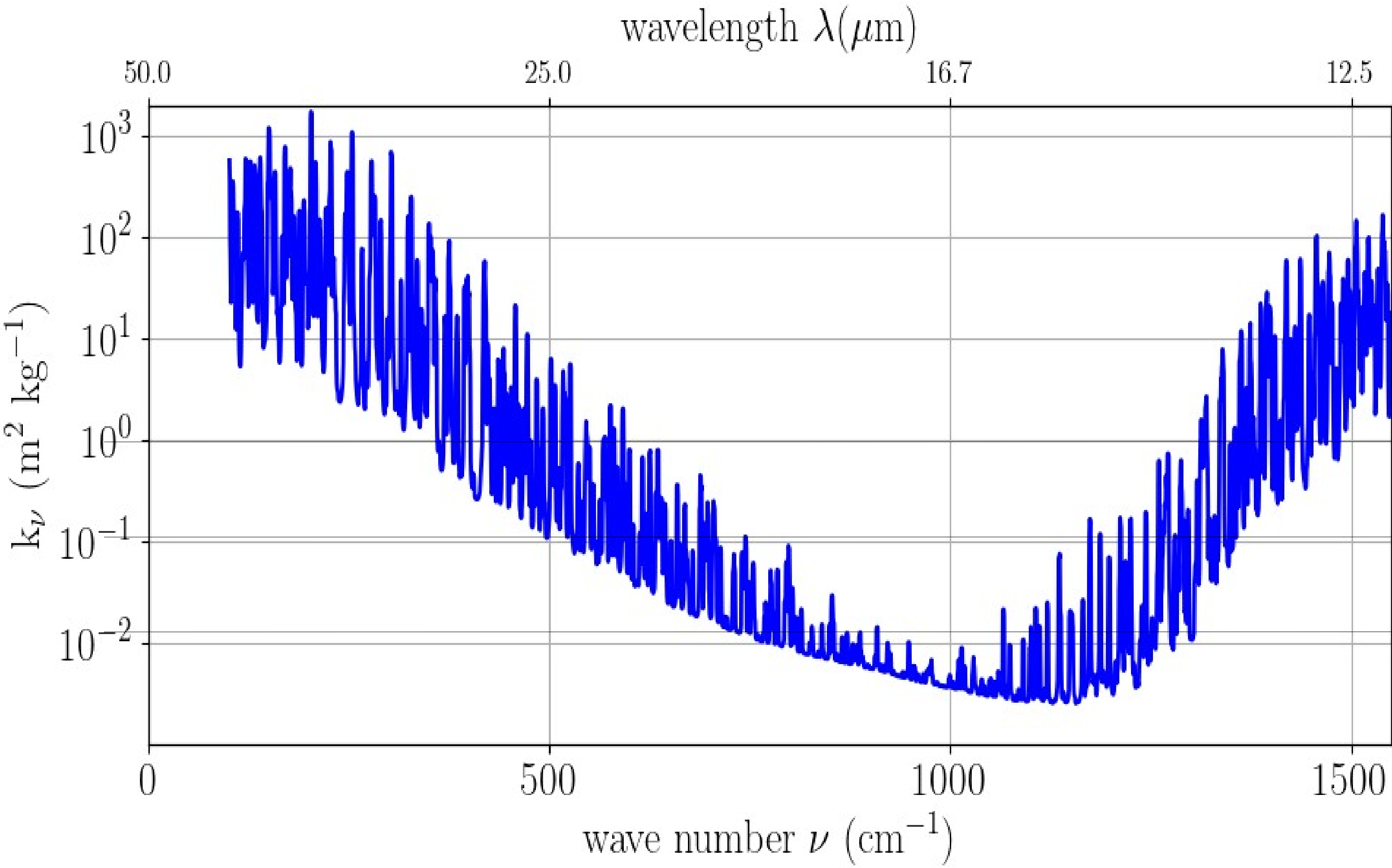
• Dans les domaines spectraux où H20 est le principal absorbant et l'atmopshère est optiquement épaisse, un changement de la température de l'atmosphère n'induit pas un changement des flux au sommet de l'atmosphère si l'humidité relative reste constante

• L'utilisation du couple température – humidité relative (T-RH) est plus pertinente celle température – humidité spécidique (T-q)

• Ces propriétés se retrouvent dans un contexte multi-modèle pour des changement du gradient vertical de température et pour un chanagement verticalement uniforme de la température

• Les incertutudes les plus importantes proviennent de l'humidité relative aussi bien en terme d'état moyen que de variation lors d'un changement climatique

optical thickness
Tau = κ * M_{H2O}



water amount for optical thickness tau=1

1 $kg.m^{-2}$,(255K)

10 $kg.m^{-2}$ (280K)

100 $kg.m^{-2}$ (310K)

Emission height

ω(z) : *conditional probability that photons emitted at z reach the TOA*

$$\omega(z) = \frac{1}{1-\Gamma_s}\frac{\partial\Gamma(z)}{\partial z}$$

Ω(z): Probability that photons *emitted by the atmosphere and that reach the TOA* ***have been emitted at altitude z***

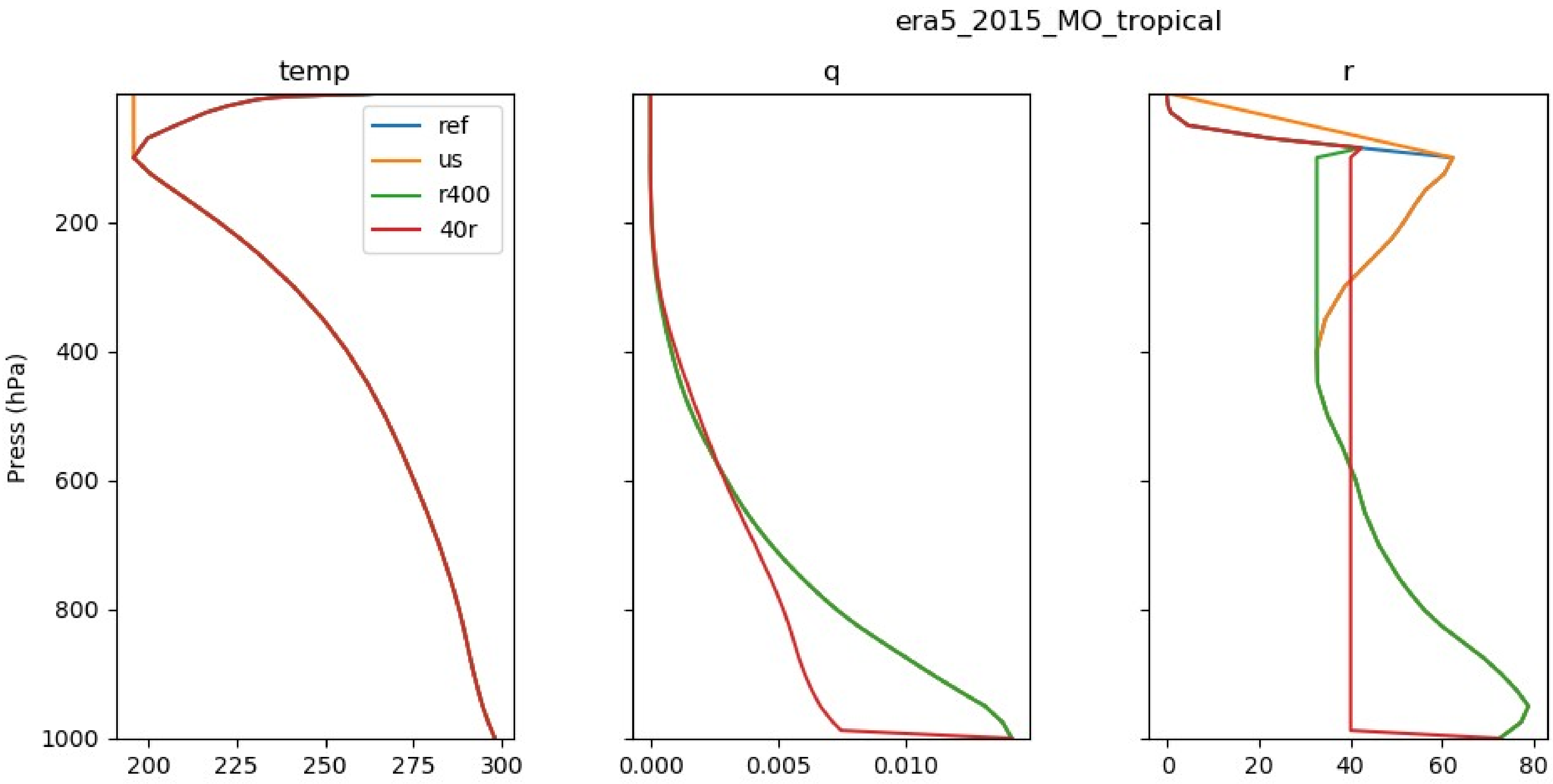
$$\Omega(z) = \frac{1}{(1-\Gamma_s)B_e}\frac{\partial\Gamma(z)}{\partial z}B(z)$$

The ***mean altitude*** *where these photons have been emitted is then:*

$$\overline{Z}_e = \int_0^H z\Omega(z)dz$$

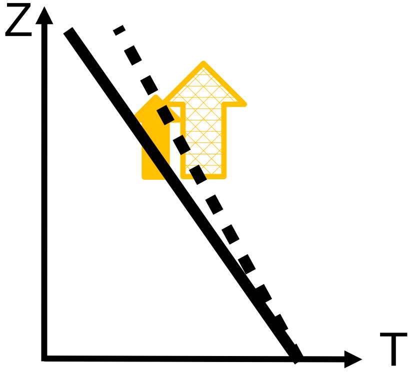
[Dufresne et al. 2020]

For a mean tropical atmosphere (ERA5)

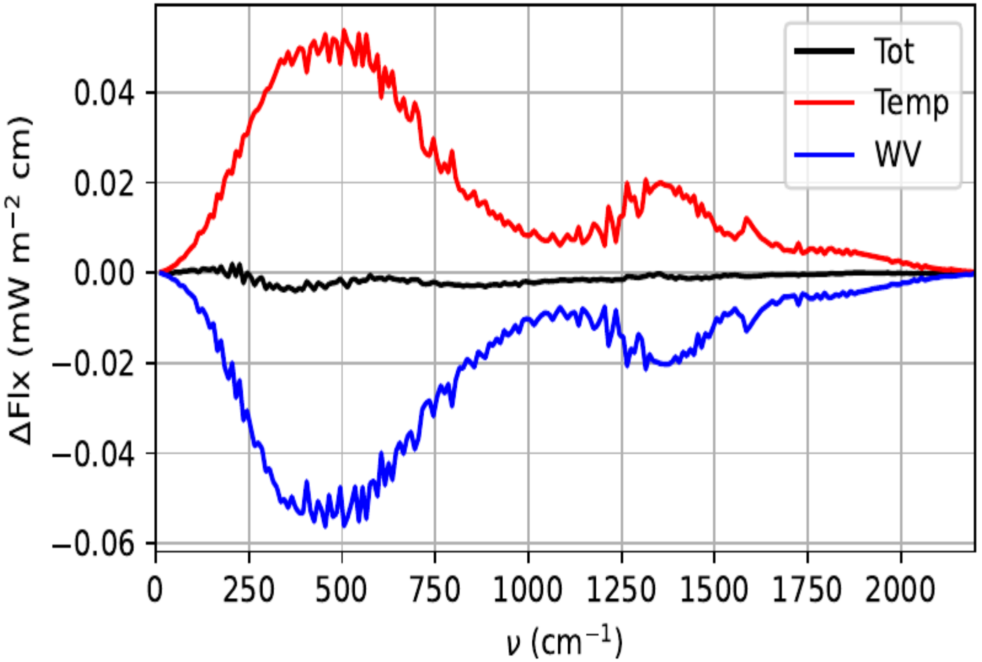


For a mean tropical atmosphere (ERA5)

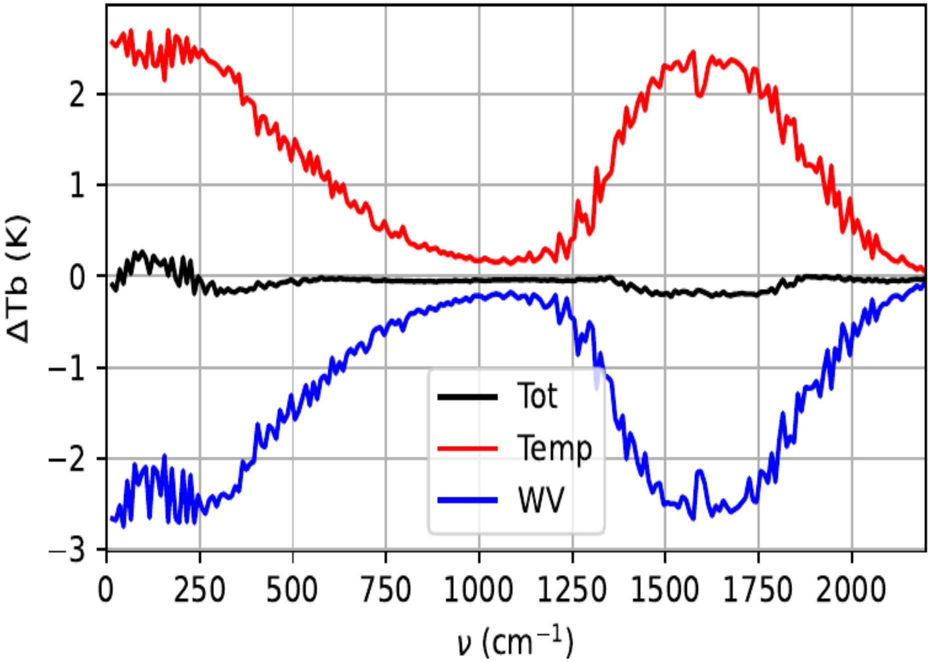
Change in brightness temperature
in response to a **change in vertical temperature gradient** (lapse rate):
0 at surface, +4K at the tropopause, H_2O only



Flux at the TOA



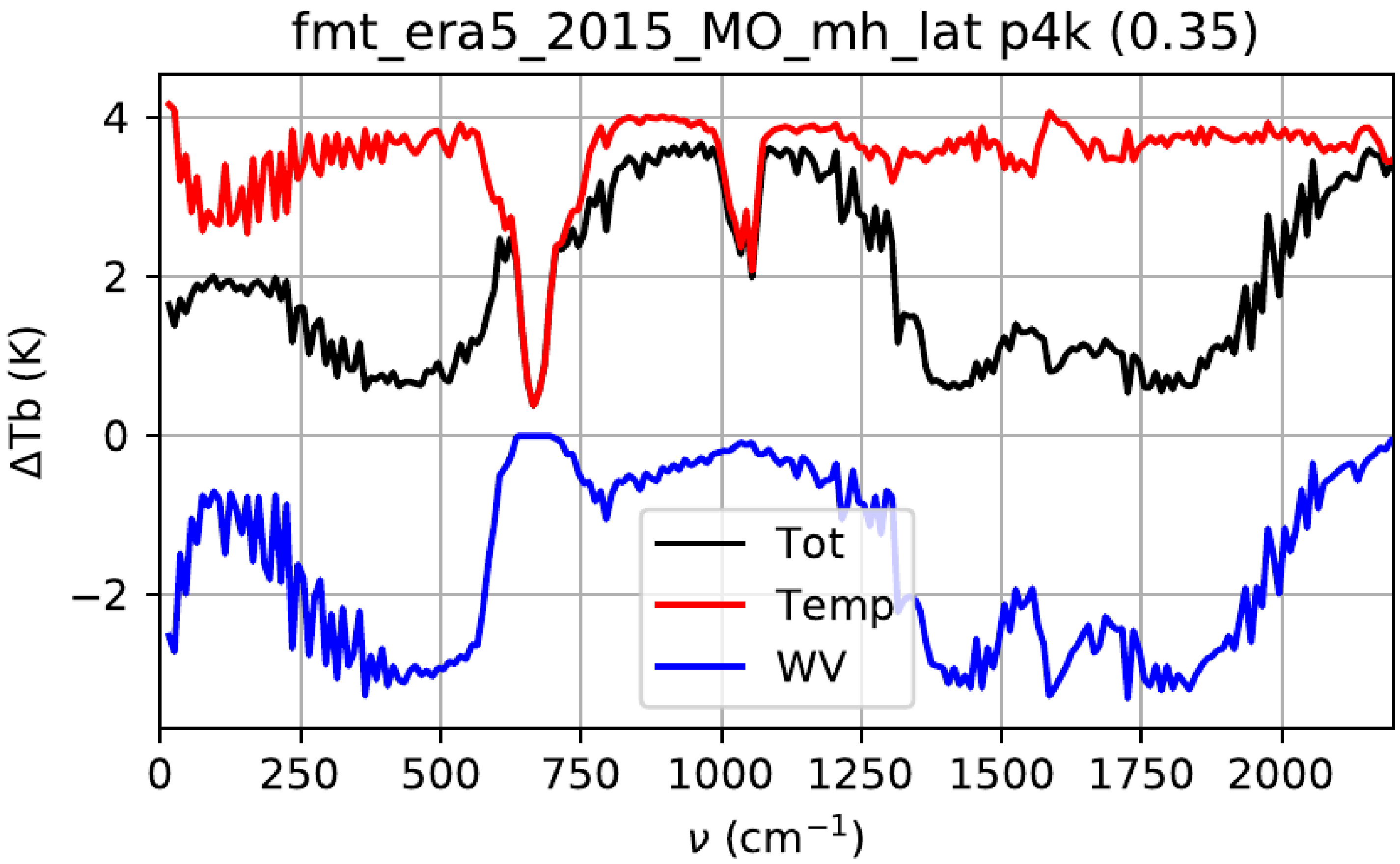
Brightness temperature



For a mean tropical atmosphere with only H_20, the **increase in water vapour to keep the relative humidity constant reduces by 107%** the increase in outgoing radiative flux due to change in lapse rate if the specific humidity would remain constant.

For mean mid-high latitude atmosphere

Change in brightness temperature
in response to a verticaly **uniform +4K increase:**



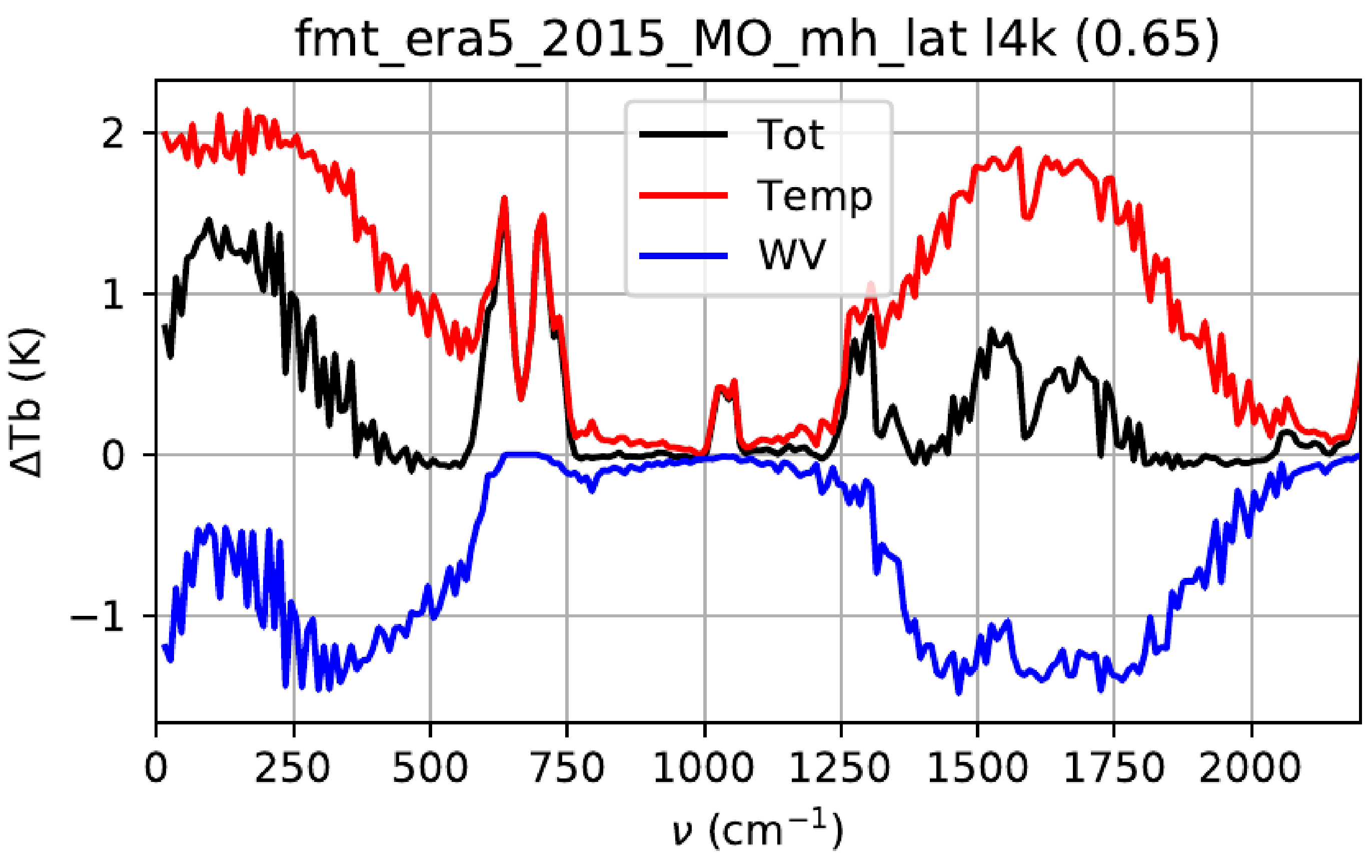
at constant **specific** humidity

at constant **relative** humidity

due to change in q **to keep relative humidity constant**

For mean mid-high latitude atmosphere

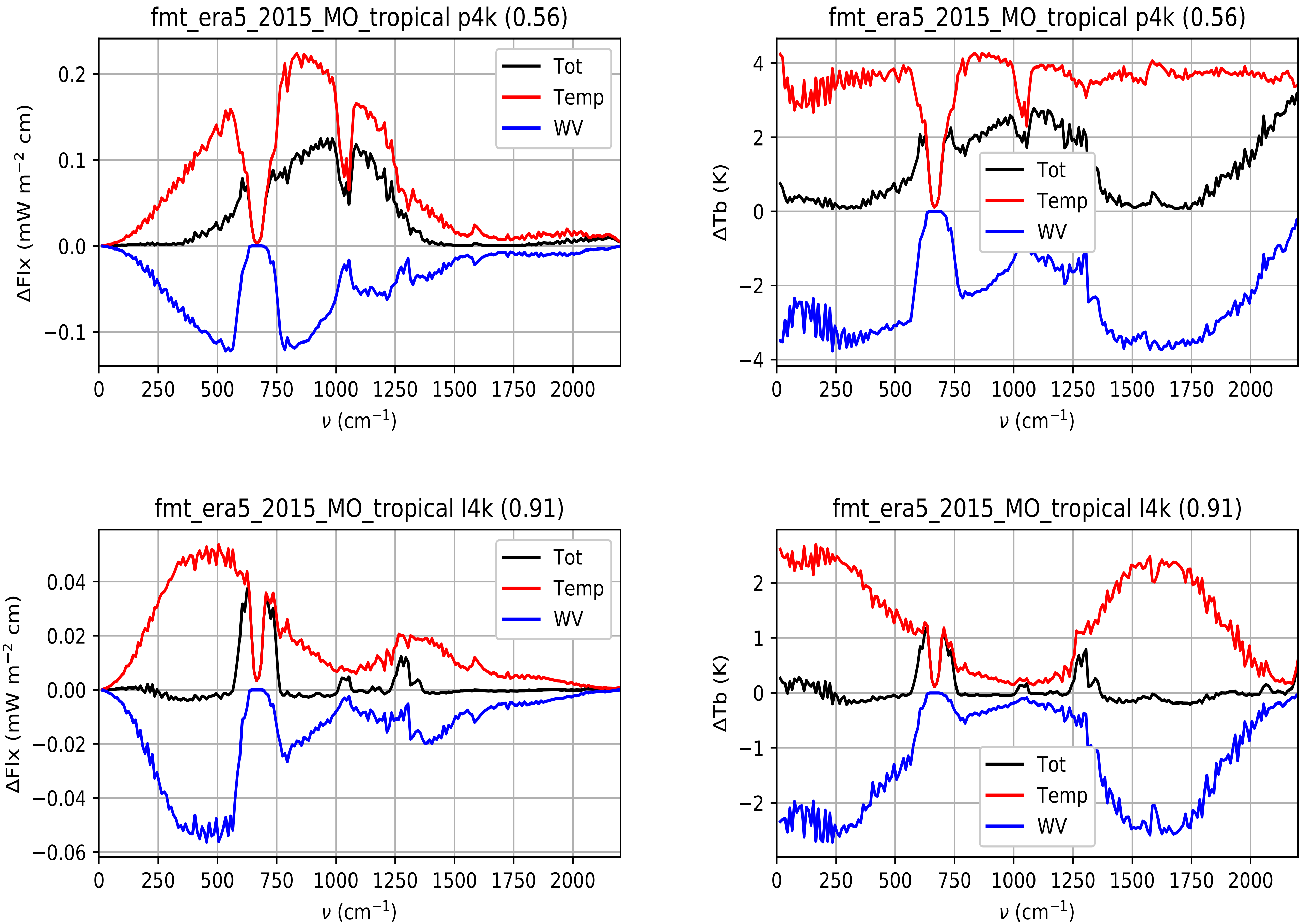
Change in brightness temperature
in response to a **change in vertical temperature gradient** (lapse rate):
0 at surface, +4K at the tropopause

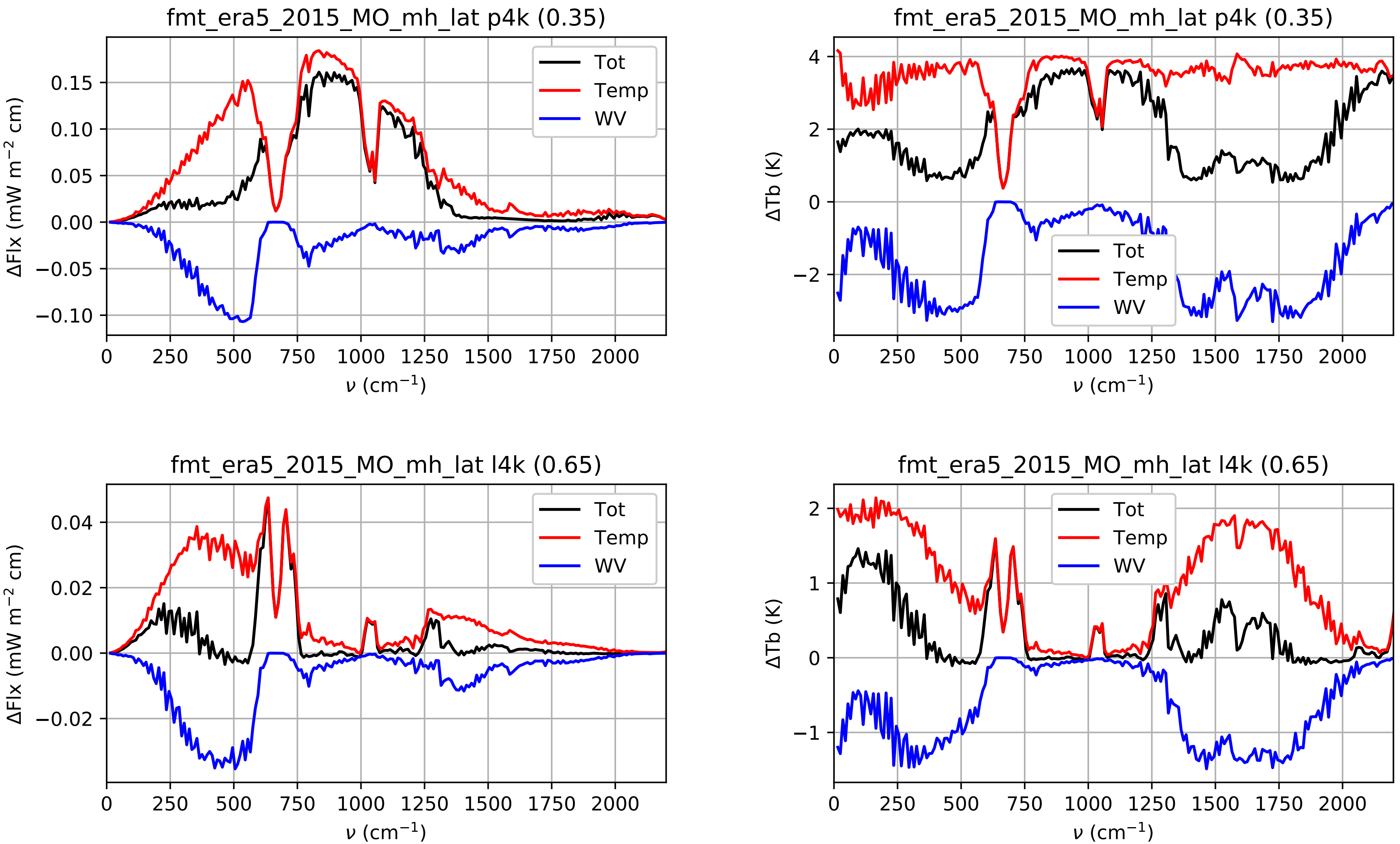


at constant **specific** humidity

at constant **relative** humidity

due to change in q **to keep relative humidity constant**

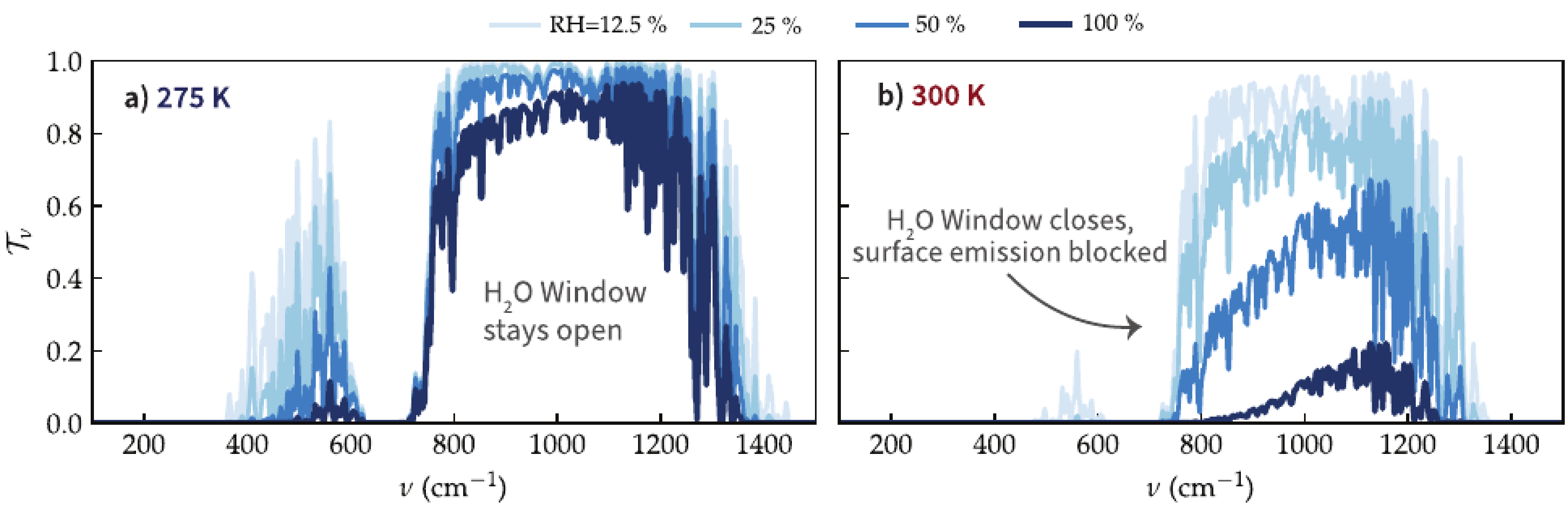

fmt_era5_2015_MO_tropical p4k (0.56)
ΔFlx (mW m^{-2} cm)
ν (cm^{-1})
Tot
Temp
WV
fmt_era5_2015_MO_tropical p4k (0.56)
ΔTb (K)
ν (cm^{-1})
Tot
Temp
WV
fmt_era5_2015_MO_tropical l4k (0.91)
ΔFlx (mW m^{-2} cm)
ν (cm^{-1})
Tot
Temp
WV
fmt_era5_2015_MO_tropical l4k (0.91)
ΔTb (K)
ν (cm^{-1})
Tot
Temp
WV



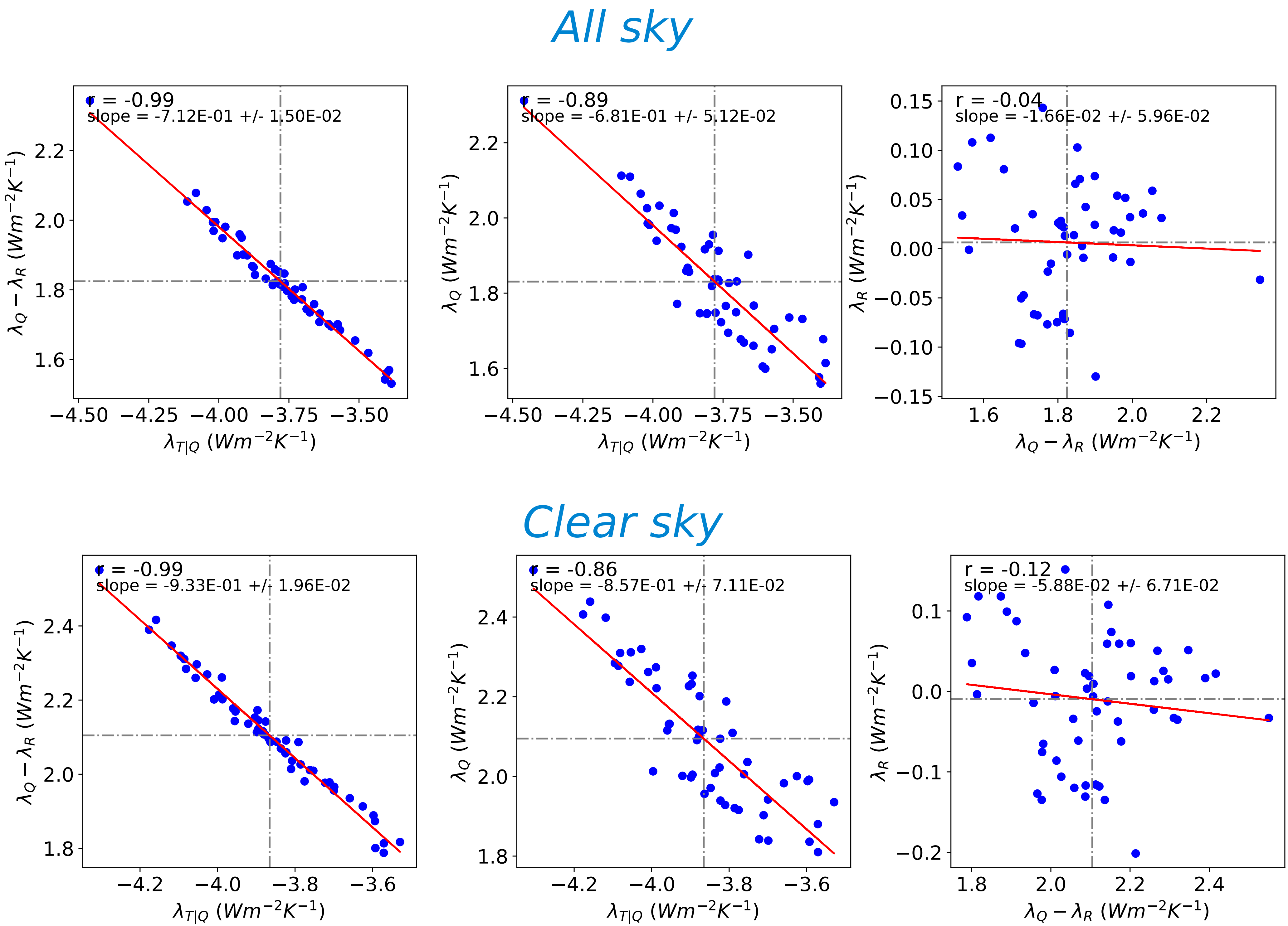
fmt_era5_2015_MO_mh_lat p4k (0.35)
ΔFlx (mW m^{-2} cm)
ΔTb (K)
fmt_era5_2015_MO_mh_lat l4k (0.65)
Tot
Temp
WV
ν (cm^{-1})

For idealized clear sky tropical atmospheres with constant relative humidity

Clear sky transmitivity



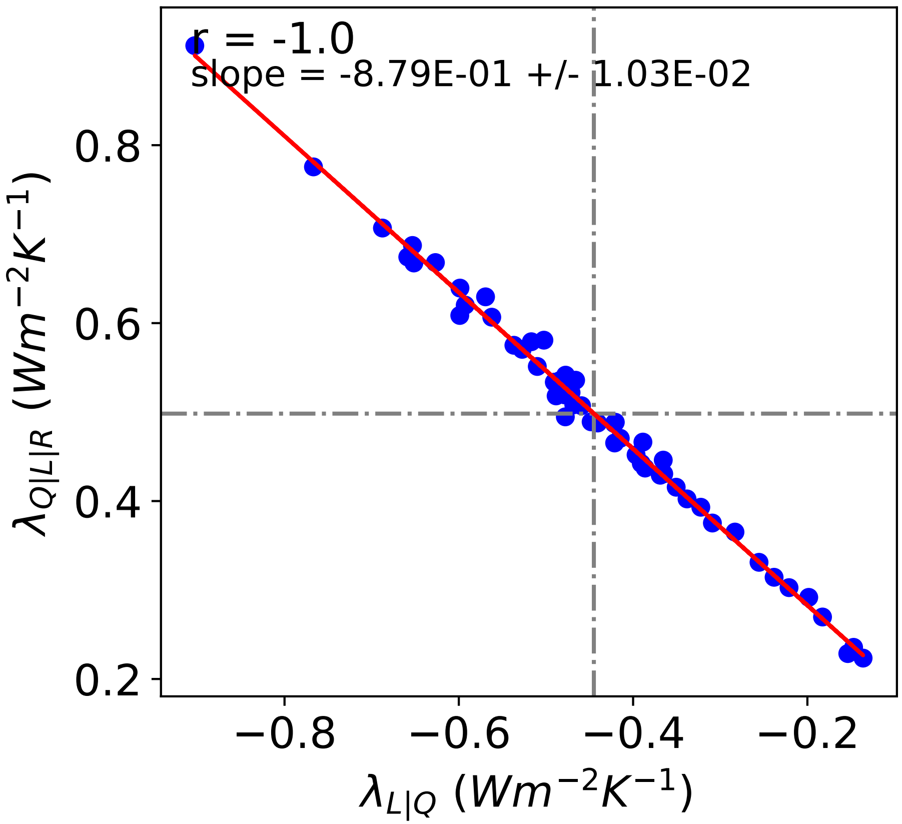
[McKim et al., GRL, smi]


All sky
r = -0.99
slope = -7.12E-01 +/- 1.50E-02
$\lambda_Q - \lambda_R$ ($Wm^{-2}K^{-1}$)
$\lambda_{T|Q}$ ($Wm^{-2}K^{-1}$)
r = -0.89
slope = -6.81E-01 +/- 5.12E-02
λ_Q ($Wm^{-2}K^{-1}$)
$\lambda_{T|Q}$ ($Wm^{-2}K^{-1}$)
r = -0.04
slope = -1.66E-02 +/- 5.96E-02
λ_R ($Wm^{-2}K^{-1}$)
$\lambda_Q - \lambda_R$ ($Wm^{-2}K^{-1}$)
Clear sky
r = -0.99
slope = -9.33E-01 +/- 1.96E-02
$\lambda_Q - \lambda_R$ ($Wm^{-2}K^{-1}$)
$\lambda_{T|Q}$ ($Wm^{-2}K^{-1}$)
r = -0.86
slope = -8.57E-01 +/- 7.11E-02
λ_Q ($Wm^{-2}K^{-1}$)
$\lambda_{T|Q}$ ($Wm^{-2}K^{-1}$)
r = -0.12
slope = -5.88E-02 +/- 6.71E-02
λ_R ($Wm^{-2}K^{-1}$)
$\lambda_Q - \lambda_R$ ($Wm^{-2}K^{-1}$)

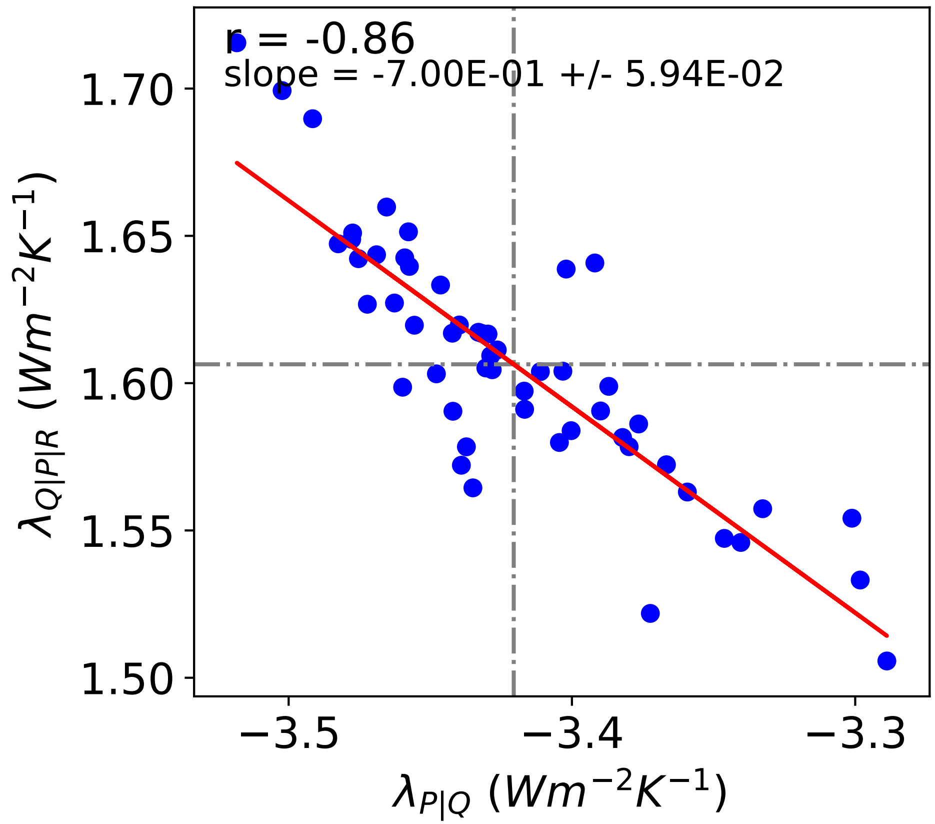
Analyse multimodèle

Clear sky, constant RH

For 50 CMIP6 models

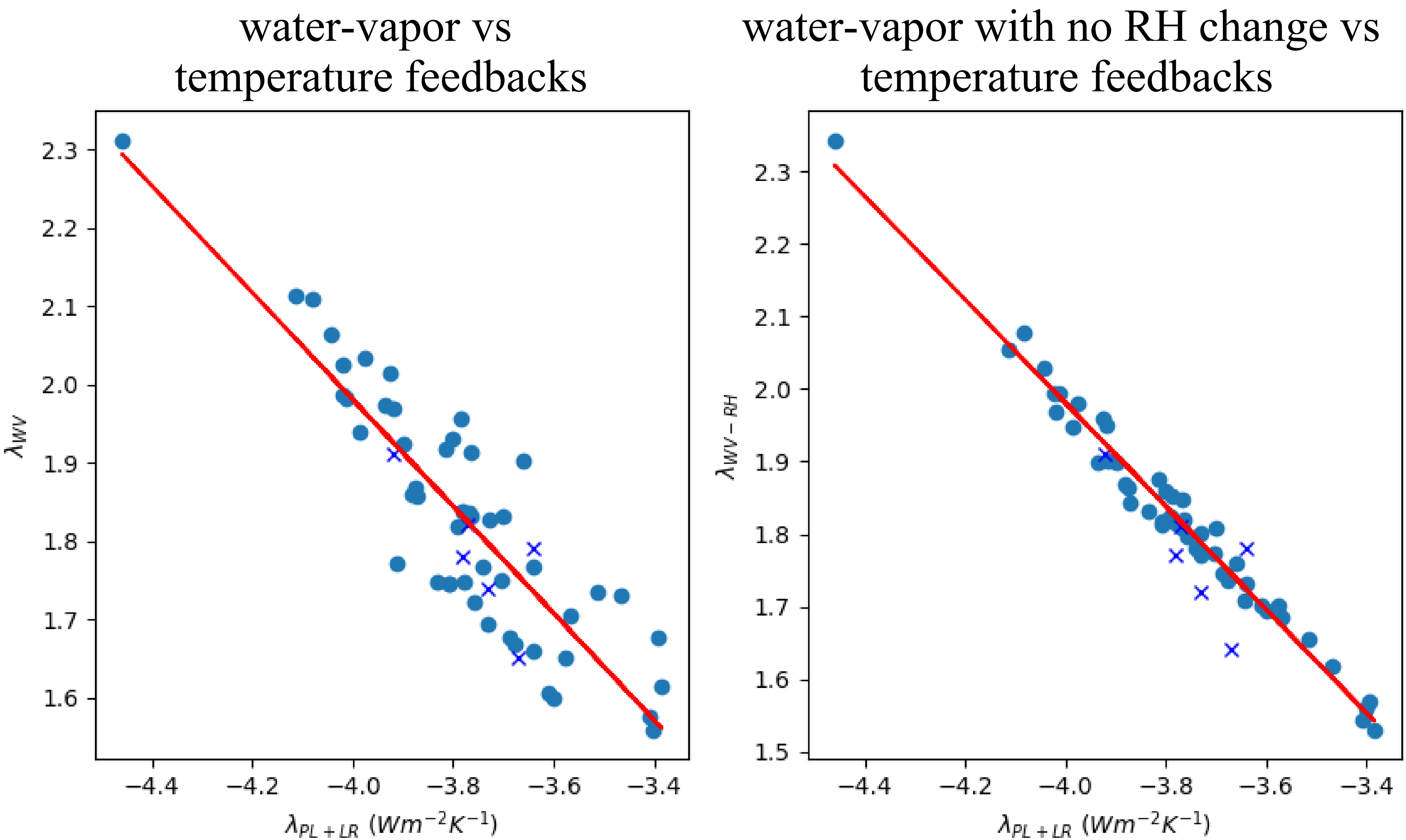


Sensibilité du flux TOA à un changement de lapse-rate à humidité constante

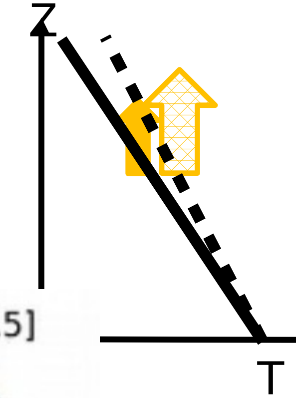


Sensibilité du flux TOA à un changement verticalement uniforme de T à humidité constante

Feedback parameters for CMIP6 models

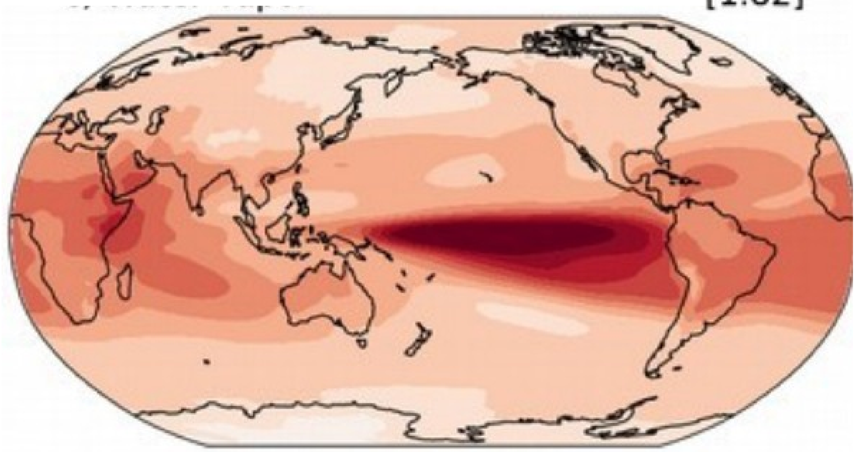


Climate feedbacks with the absolute and relative humidity decompositions

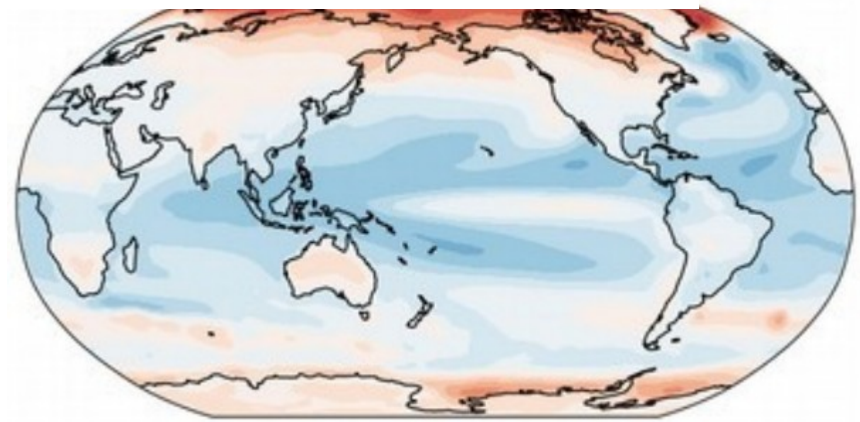


due to change in temp. vert. gradient

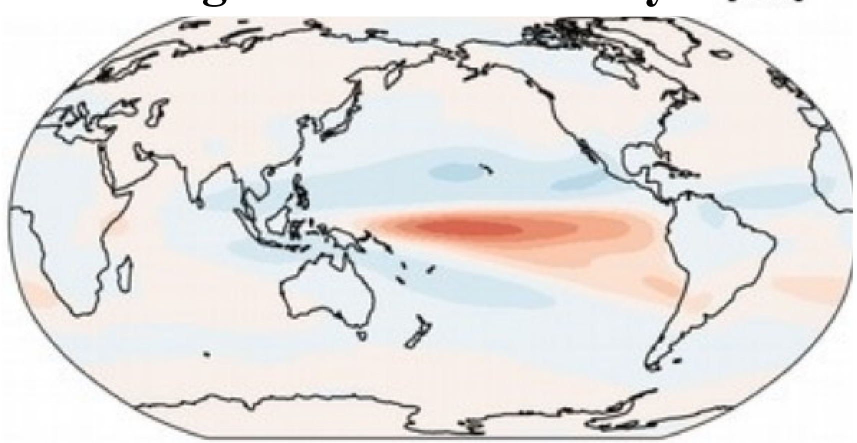
due to change in specific humity (kg/kg) [1.82]



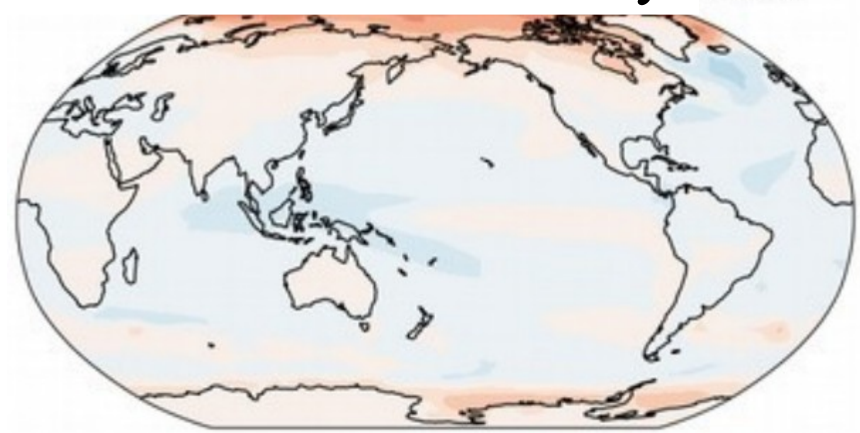
at consant specific humity [-0.5]

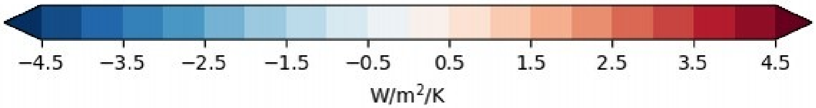


due to change in relative humity [0.0]



at consant relative humity -0.05]

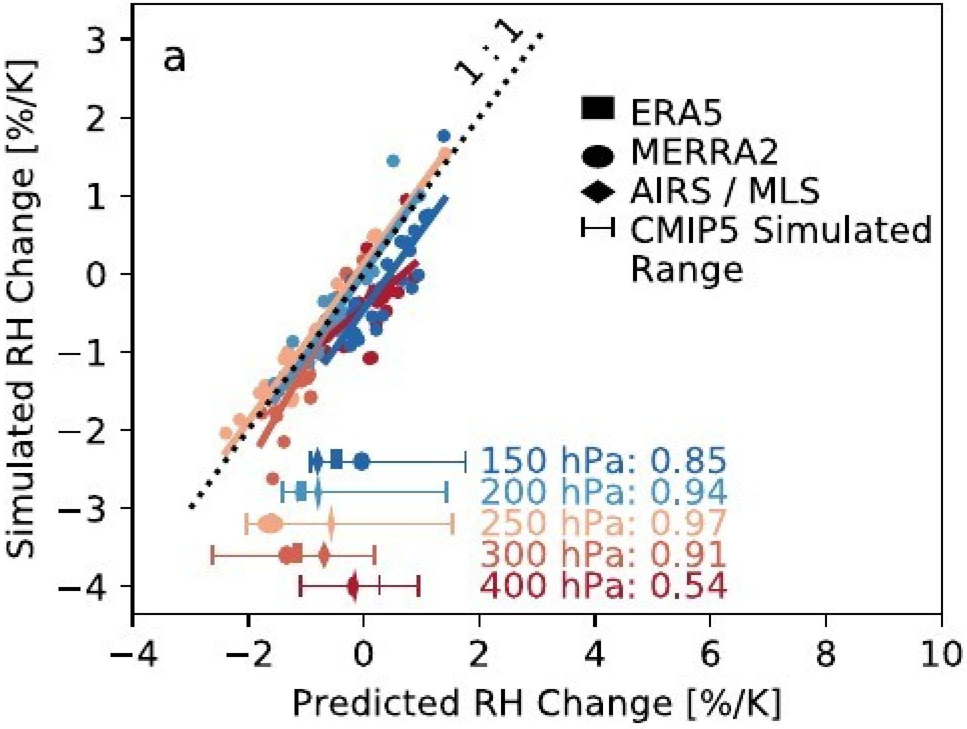


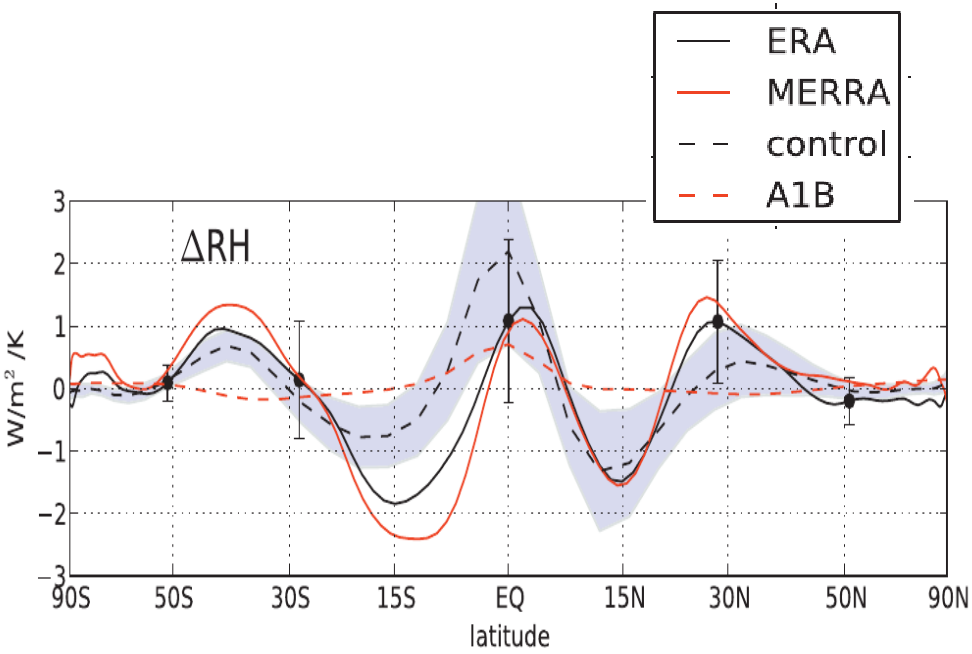


[courtesy of M. Zelinka 2021]
(https://doi.org/10.5281/zenodo.5206851)

Water vapour feedback

Little change of the relative humidity with temperature at inter-annual time scale

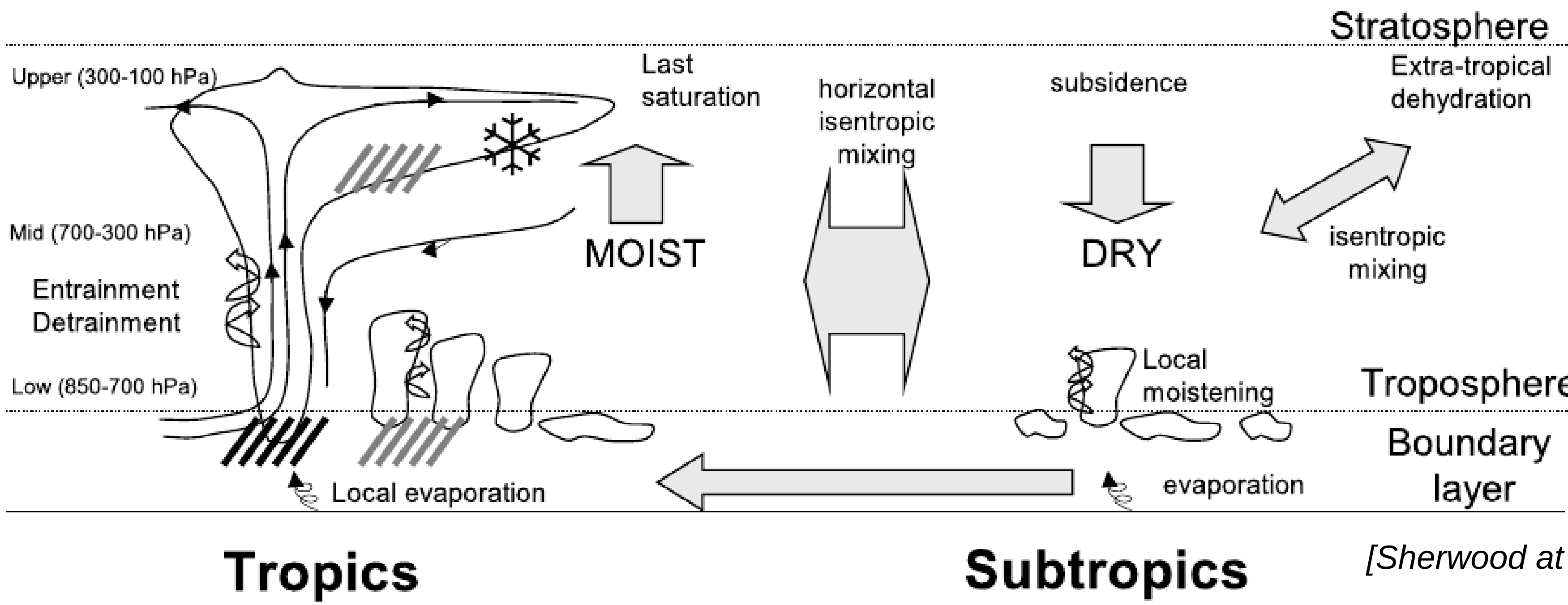




[Dessler etal., 2013]

[Po-Chedley et al., GRL,2019]

Water vapour feedback : The last saturation paradigm

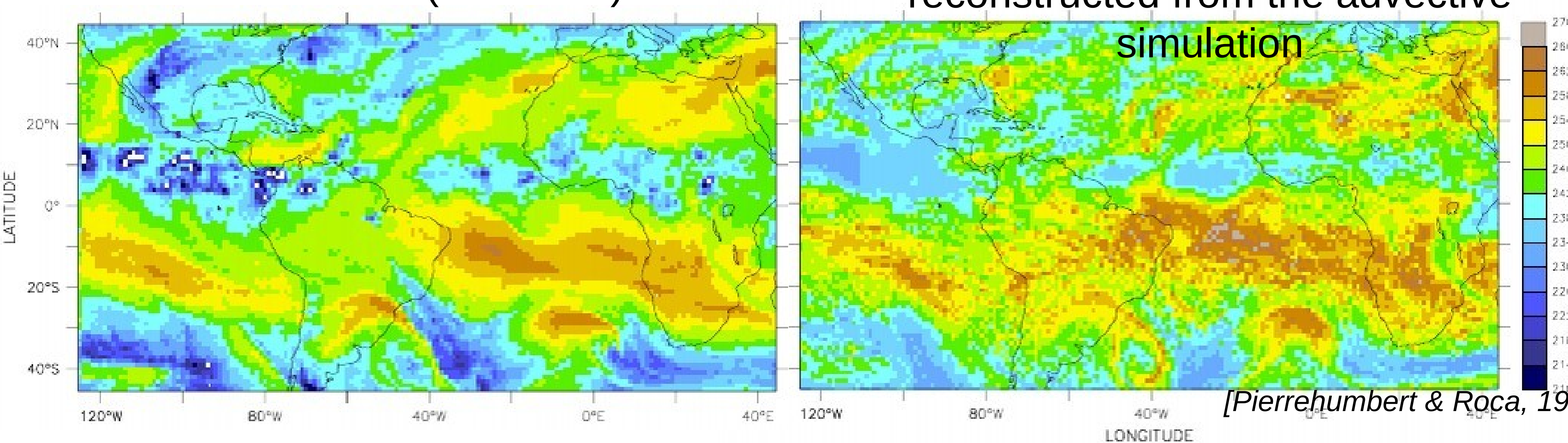


[Sherwood at al., 2010]

6.3 µm brightness temperature (~ relative humidity)

Observed (Meteosat)

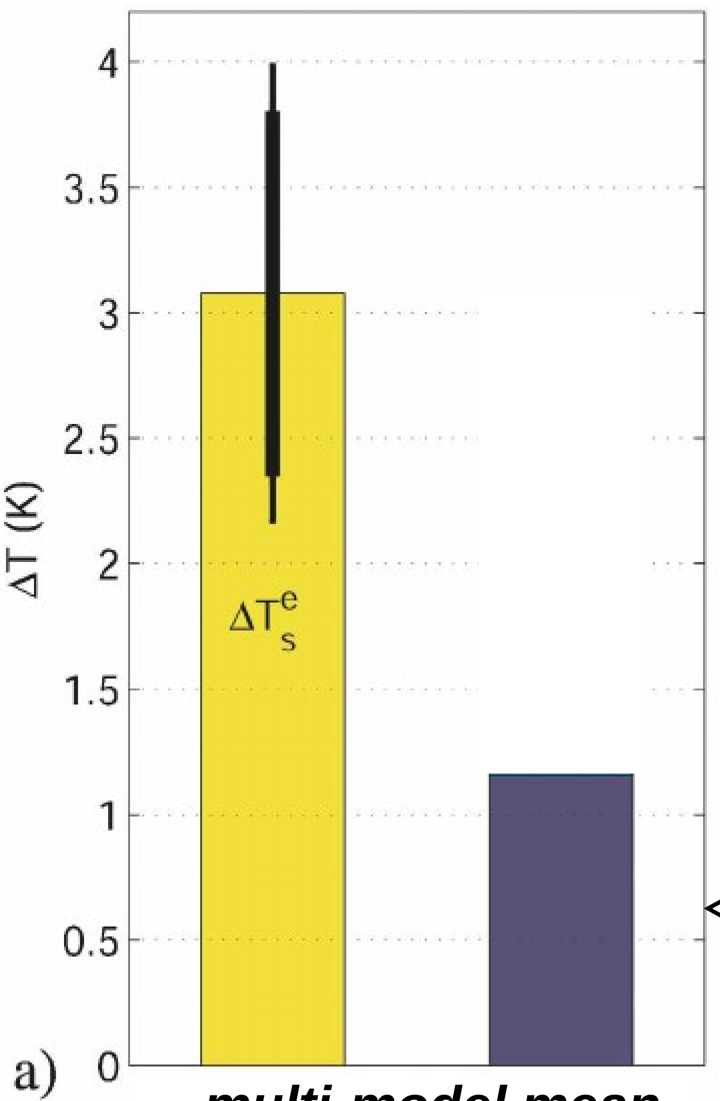
reconstructed from the advective simulation



[Pierrehumbert & Roca, 1998]

How much individual feedbacks contribute to global warming

Equilibrium temperature response to a CO_2 doubling



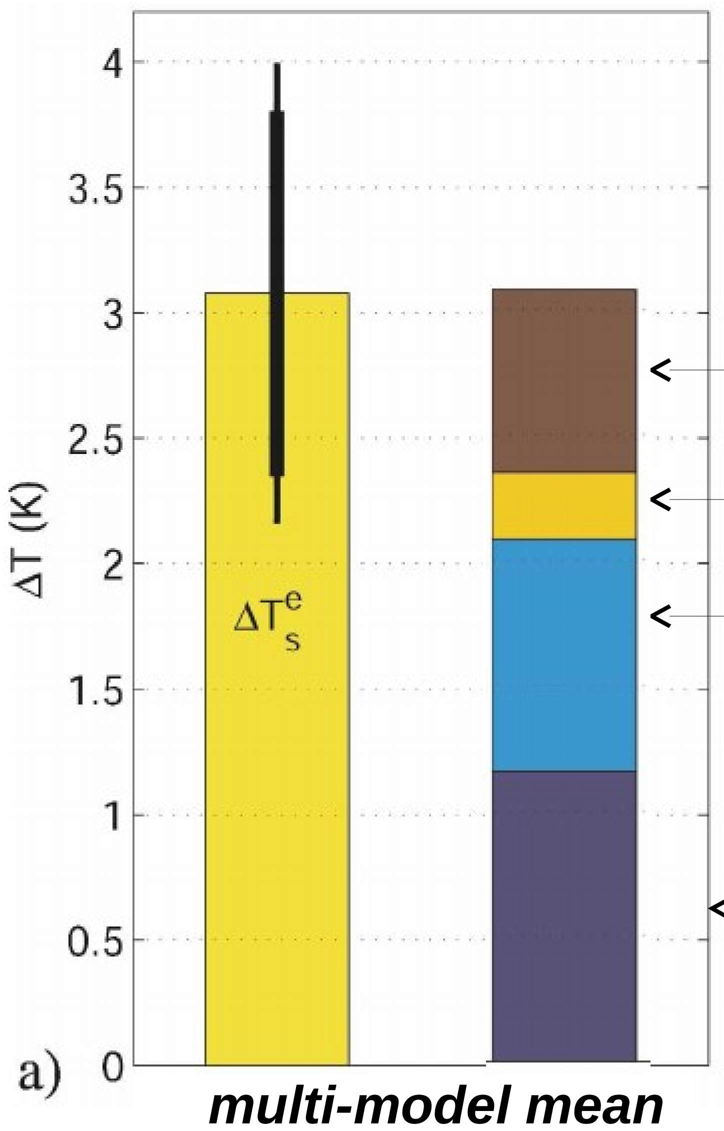
multi-model mean

Direct response to the forcing
Planck response

(Dufresne & Bony, 2008)

How much individual feedbacks contribute to global warming

Equilibrium temperature response to a CO_2 doubling



Climate feedbacks: Indirect response to the forcing

clouds

snow and ice (surface albedo)

water vapor

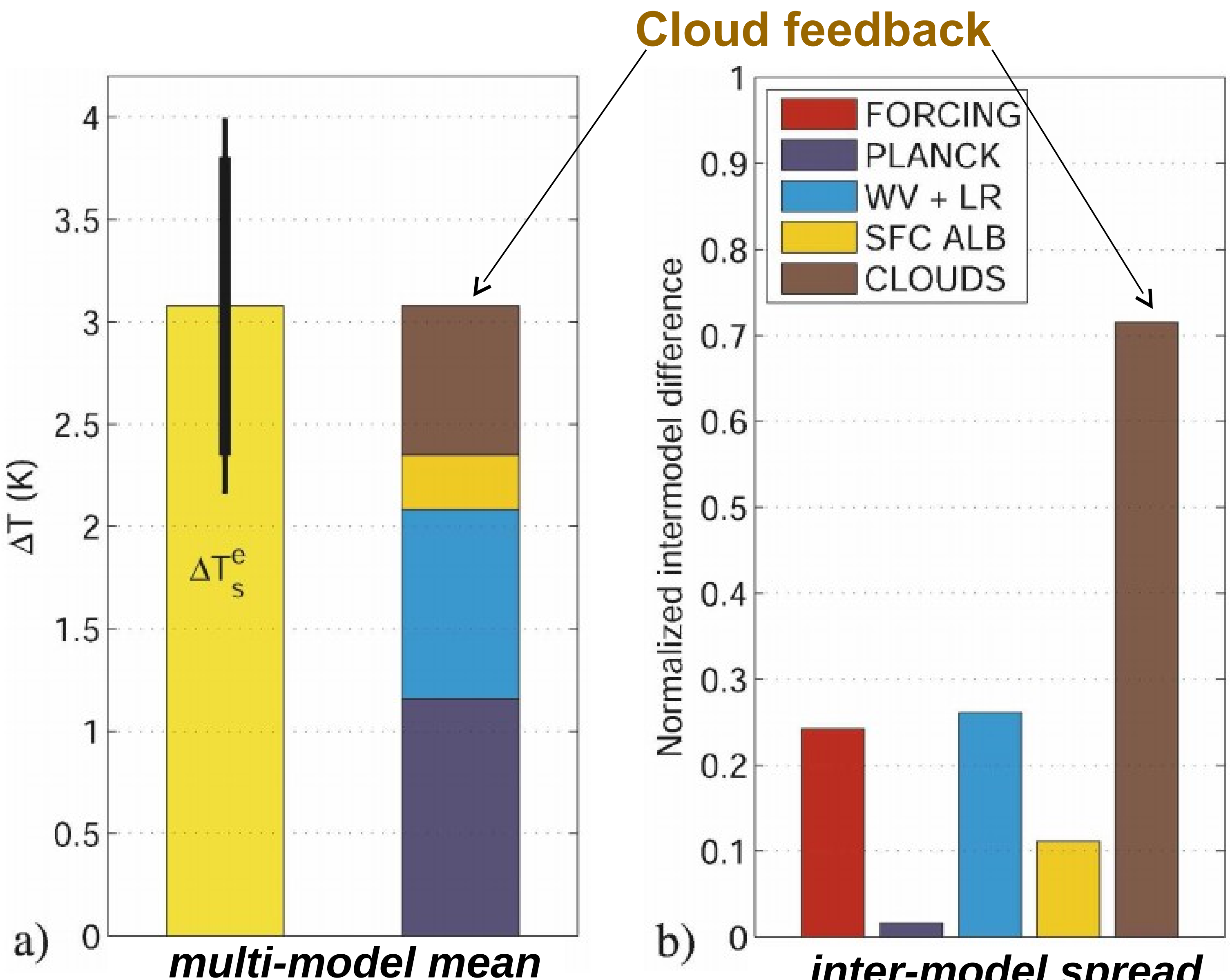
Direct response to the forcing

Planck response

(Dufresne & Bony, 2008)

How much individual feedbacks contribute to global warming

Equilibrium temperature response to a CO_2 doubling



(Dufresne & Bony, 2008)

How to compute feedbacks ?

Diagnostic of feedback parameters through the Kernel approach

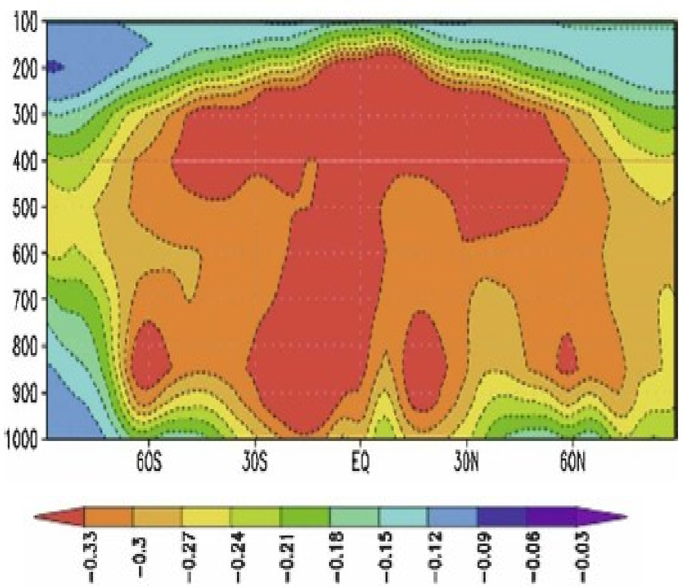
$$\lambda = \frac{dN}{dT_s} = \sum_x \frac{\partial N}{\partial x} \frac{\partial x}{\partial T_s}$$

radiative kernel computed by radiative codes ($\frac{\partial N}{\partial x}$)

response to surface temperature change ($\frac{\partial x}{\partial T_s}$)

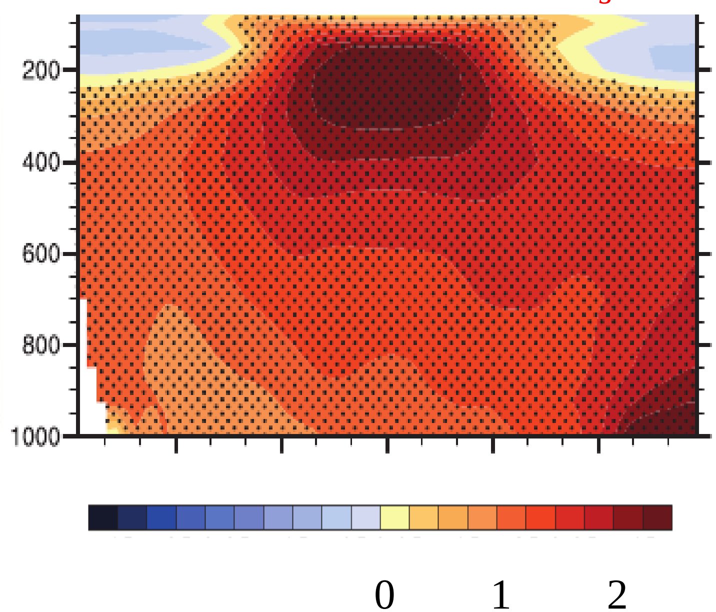
e.g. for *x* = *T* :

Temperature kernel $\frac{\partial N}{\partial T}$



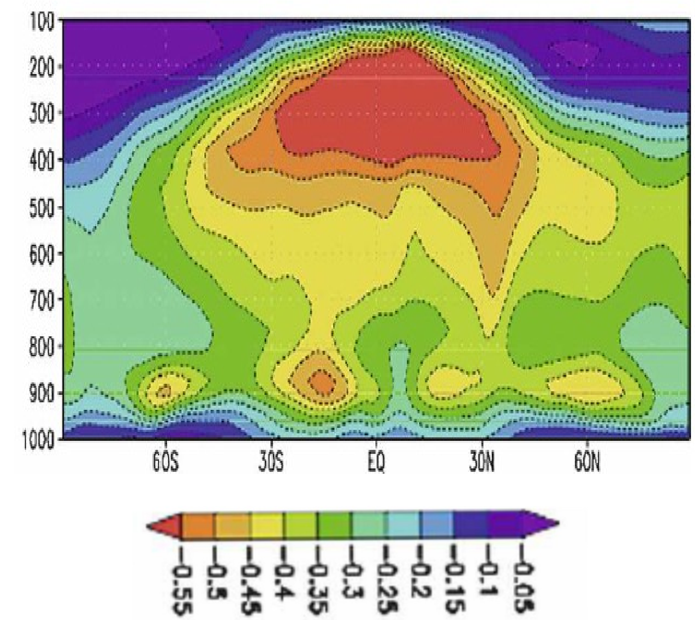
W/m²/K/(100hPa)

Temperature change $\frac{\partial T}{\partial T_s}$



K/K (approximate)

Temperature feedback parameter $\lambda_T = \frac{\partial N}{\partial T} \frac{\partial T}{\partial T_s}$



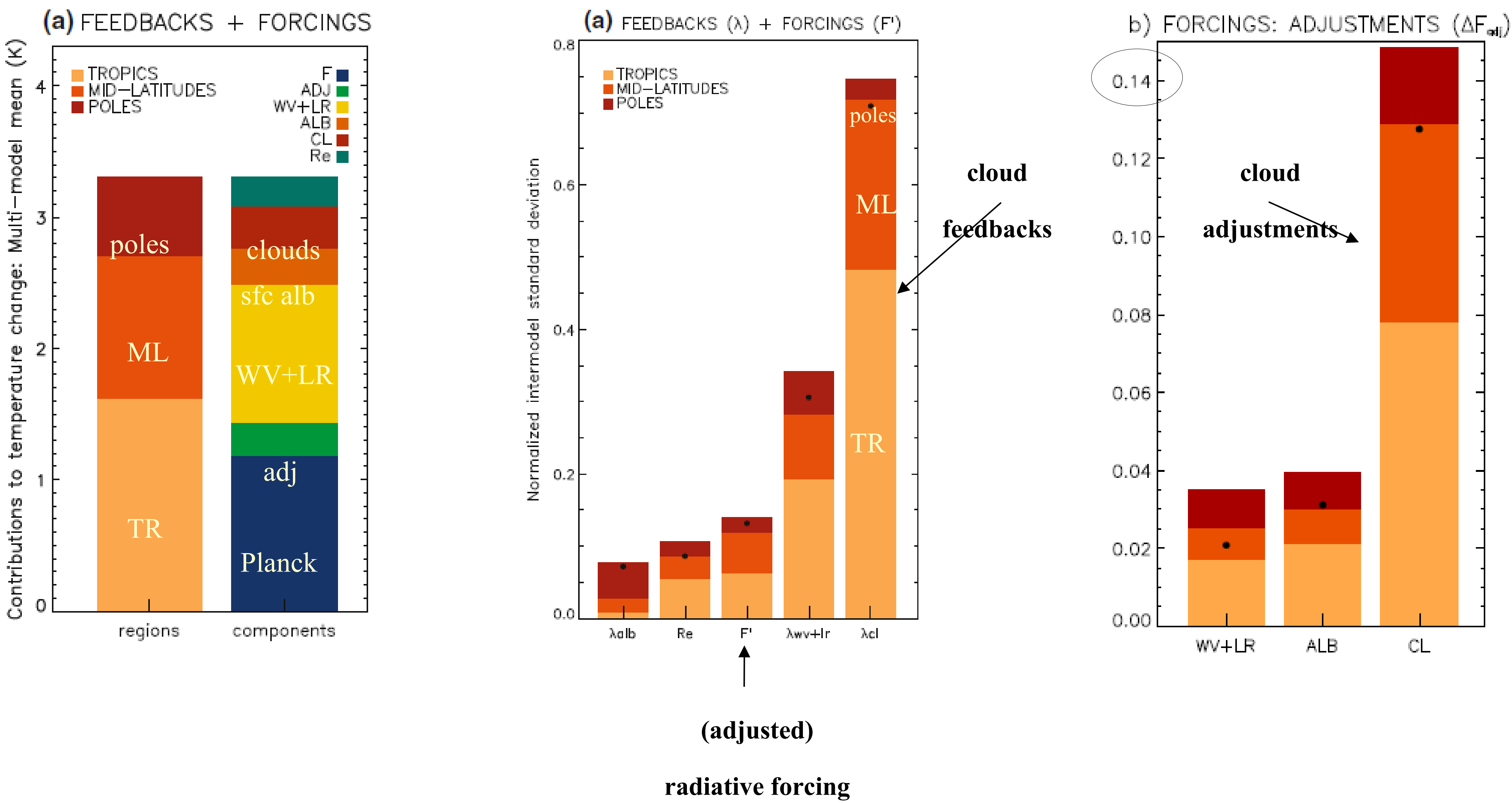
W/m²/K/(100hPa)

Soden et al., J. Climate, 2008

Decomposition of CMIP5 climate sensitivity estimates

Multi-Model Mean

Inter-Model Spread



Cloud feedbacks still constitute a leading source of uncertainty.

Vial et al., Clim. Dyn., 2013

Revised Framework

let's assume that $R = R(\varphi, T_s, X)$ $\quad X = X(\varphi, T_s)$

$$\Delta R = \left[\left(\frac{\partial R}{\partial \varphi} \right)_{T_s, X} + \left(\frac{\partial R}{\partial X} \right)_{\varphi, T_s} \frac{\partial X}{\partial \varphi} \right] \Delta \varphi + \left[\left(\frac{\partial R}{\partial T_s} \right)_{\varphi, X} + \left(\frac{\partial R}{\partial X} \right)_{\varphi, T_s} \frac{\partial X}{\partial T_s} \right] \Delta T_s$$

instantaneous radiative change + *adjustments to φ*

Planck response + *radiative feedbacks*

radiative forcing

(named « effective radiative forcing » in AR5)

climate response

Water vapor + lapse rate feedbacks

RH ≈ constant
Clausius Clapeyron

increase of WV

increase of greenhouse effect

decrease of OLR

positive feedback

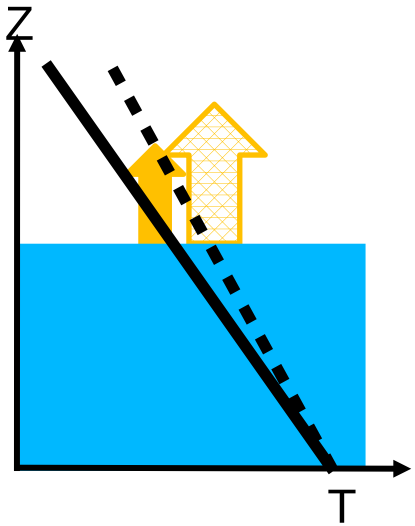
theory

decrease of moist adiabatic lapse rate

decrease of lapse rate

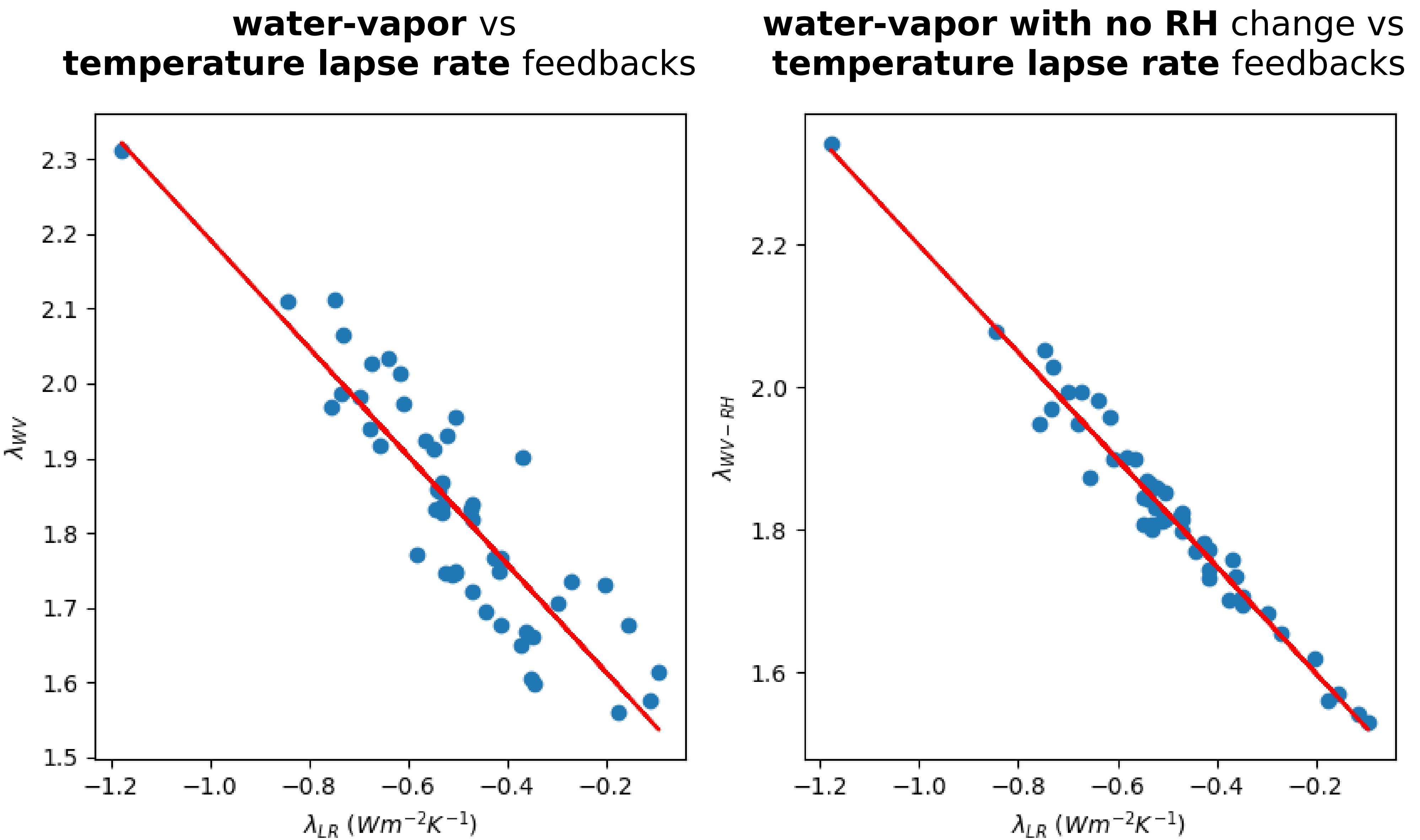
increase of OLR

negative feedback

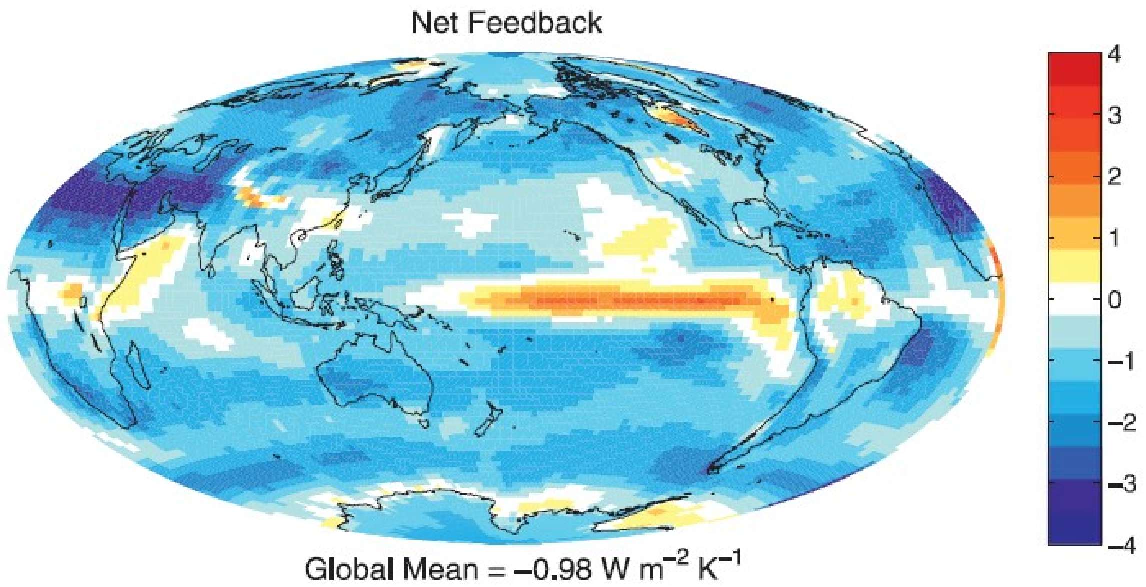


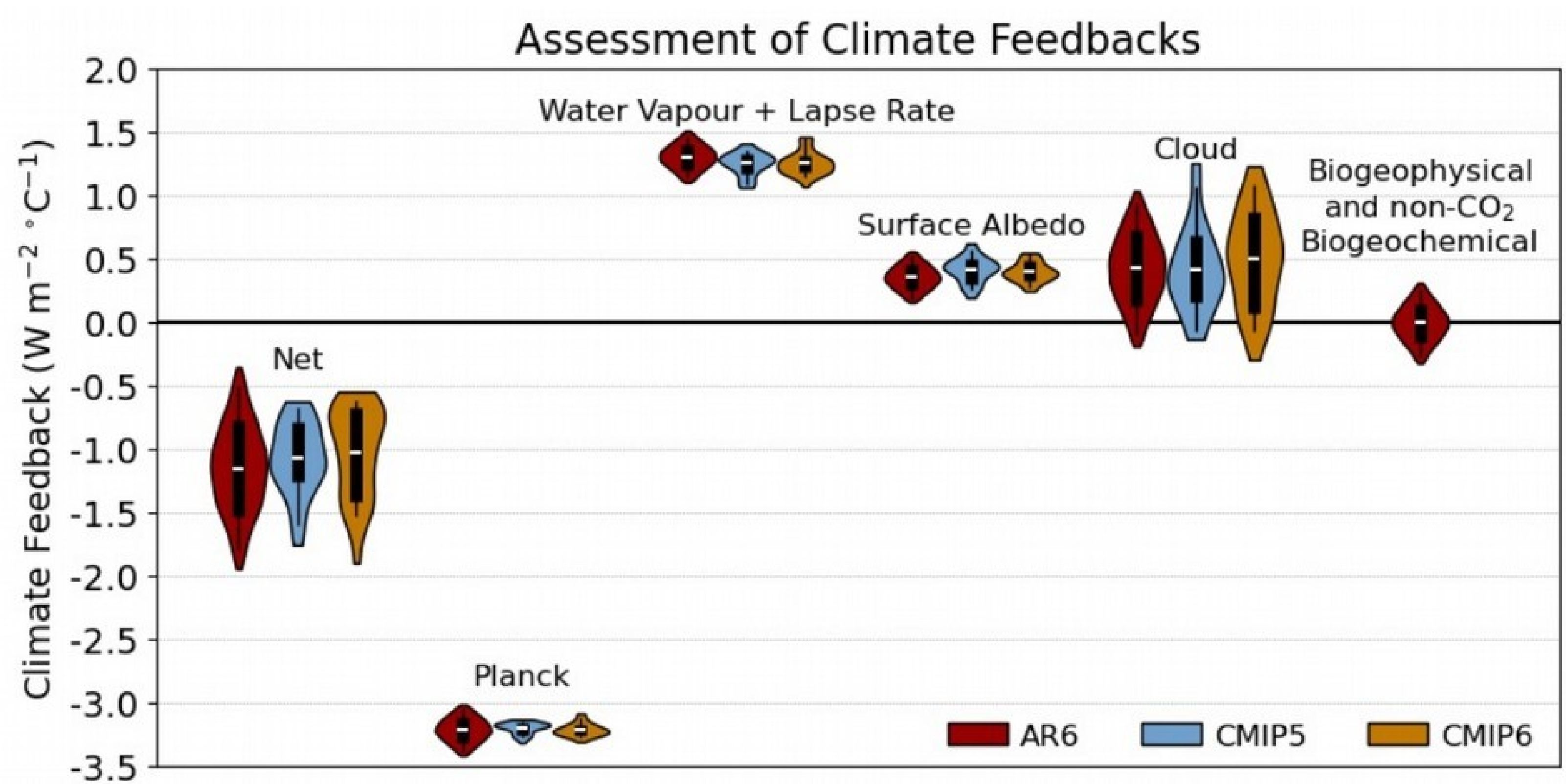
positive feedback

Feedback parameters for CMIP6 models



Climate total feedback





[IPCC AR6 WG1, ch 7]