The saturated greenhouse effect theory of Ferenc Miskolczi

Presented by

Miklós Zágoni

The Miskolczi-principle

- **The greenhouse effect is not a free variable.**
- Earth type planetary atmospheres, having partial cloud cover and sufficient water vapor reservoirs, maintain an ***energetically maximized*** (constant, ‘saturated’) greenhouse effect that cannot be increased by emissions.
- The following presentation serves the proof of the above statements.

Content

I. The tool: HARTCODE
II. Applications on the TIGR database
III. Results of computations
IV. Consequences
V. Historical perspective
VI. Interpretation and comments
VII. Further considerations

Greenhouse temperature

- The greenhouse effect is the difference between the global average yearly surface temperature (+15C, 288K) and the „effective" temperature (–18C, 255K):
- $T_g = T_s - T_e = 288 - 255 = 33°C$
- This is the amount of surplus temperature, coming from the presence of clouds and „greenhouse" gases in the atmosphere (mainly H_2O water vapor and CO_2 carbon dioxide).

Greenhouse factors and functions

- The greenhouse factor (**G**) is the difference between the surface upward longwave radiation (**S_U**) and the outgoing longwave radiation (**OLR**) (Raval and Ramanathan, 1989):
  $G = S_U – OLR$ $(S_U=\sigma T_s^4;\ OLR=\sigma T_e^4)$.
- The **g** normalized greenhouse factor (or greenhouse function) is **$g = G / S_U$** .
- The **f** transfer function can be defined as
  $f = OLR / S_U$ $(f = 1 – g)$.
- If we want to relate **G** (or **g**, or **f**) to the amount of the atmospheric IR absorbers,
- we have to use **radiative transfer codes**.

Wikipedia: public line-by-line radiative transfer codes

GENLN2 [[9]]	Edwards (1992); Edwards (1987)		Edwards line-by-line atmospheric transmission/radiance model
LinePak	GATS		Transmittance/radiance modeling software
LBLRTM [[10]], [[11]]	Atmospheric and Environmental Research		Line By Line Radiative Transfer Model

FASCODE is another notable line by line code which uses HITRAN.

Part I. HARTCODE
Description and validation

- **High Resolution Atmospheric Radiative Transfer LBL Code (Miskolczi et al., 1989)**
  http://miskolczi.webs.com/**hartcode_v01.pdf**
- **Verified against GENLN2, LinePak, LBLRTM and FASCODE:**

  Kratz-Mlynczak-Mertens-Brindley-Gordley-Torres-Miskolczi-Turner: An inter-comparison of far-infrared line-by-line radiative transfer models. **Journal of Quantitative Spectroscopy & Radiative Transfer No. 90, 2005**.
- F. Miskolczi and R. Guzzi: Effect of nonuniform spectral dome transmittance on the accuracy of infrared radiation measurements using shielded pyrradiometers and pyrgeometers. **Applied Optics, Vol. 32. No. 18., 1993**.
- Rizzi-Matricardi-Miskolczi: Simulation of uplooking and downlooking high-resolution radiance spectra with two different radiative transfer models. **Applied Optics, Vol. 41. No. 6, 2002**.

Description

- HARTCODE computes the spectral and integrated optical depth, transmittance and radiance flux densities of a layered atmosphere for any viewing geometry with the desired resolution using the line-by-line integration method. The ultimate spectral resolution is not limited and – due to numerical reasons – the accuracy of the channel transmittances/radiances is more than six significant figures.
- The atmosphere up to 61 km was stratified using 32 exponentially placed layers with about 100 m and 10 km thickness at the bottom and the top. The directional radiances were determined in nine streams. Assuming cylindrical symmetry, they were integrated over the hemispheric solid angle. The upward and downward slant paths were identical, assuring that the directional spectral transmittances for the reverse trajectories were equal.

- Eleven absorbing IR active molecular species were involved: H_2O, CO_2, O_3, N_2O, CH_4, NO, SO_2, NO_2, CCl_4, F11 and F12.
- Inputs: observed or model-based thermodynamic and spectroscopic atmosphere-databases.
- HARTCODE's direct outputs are: spectral surface upward radiation ($S_{\nu,U}$), downward atmospheric emittance ($E_{\nu,D}$), upward atmospheric emittance ($E_{\nu,U}$), and transmitted surface radiance ($S_{\nu,T}$). In this way, HARTCODE is able to partition the spectral outgoing longwave radiation (OLR_ν) into $S_{\nu,T}$ and $E_{\nu,U}$ ($OLR_\nu = S_{\nu,T} + E_{\nu,U}$). The spectrally integrated quantities are indicated by omitting the subscript ν .

The Far-Infrared Spectrum

Exploring a New Frontier in the Remote Sensing of Earth's Climate

A White Paper Submitted to the National Research Council

May 16, 2005

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3.2 Water Vapor Greenhouse Effects and Radiative Cooling of the Atmosphere

In order to assess the variability of the far-infrared at the top of the atmosphere, *Miskolczi and Mlynczak* [2004] computed the clear sky top-of-atmosphere fluxes in three spectral bins, the far-infrared, the mid-infrared, and window regions. The window region covered 720 to 1260 wavenumbers, and the mid-infrared covers the difference between far-infrared and window regions to 3000 wavenumbers.

Spectral clear-sky OLR

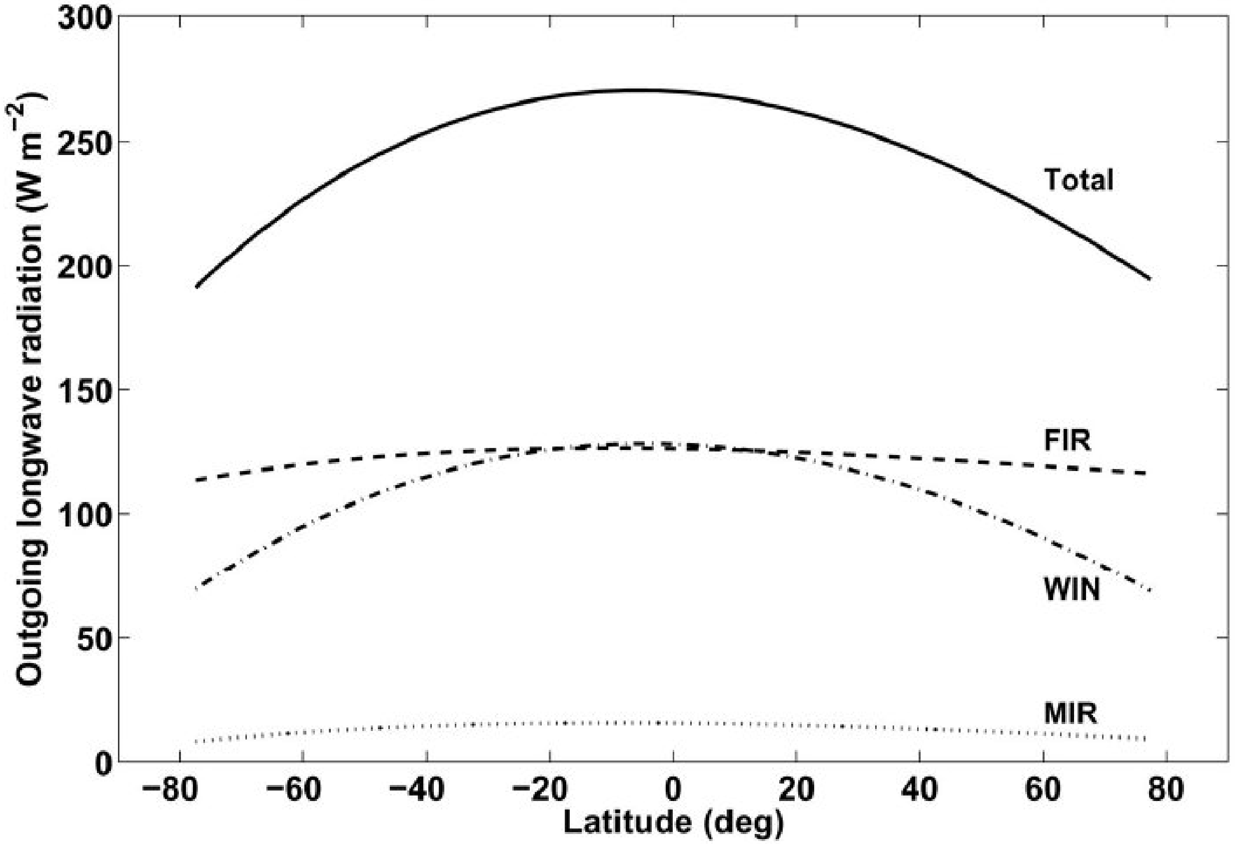


Figure 3. Spectral decomposition of the clear-sky outgoing longwave radiation (OLR). The curves are labeled for the total OLR, and that due to the far-infrared (FIR), window (WIN), and mid-infrared (MIR).

surface (i.e., the greenhouse effect). The distribution of water vapor and associated radiative forcings and feedbacks are well recognized as major uncertainties in understanding and predicting future climate [*Lindzen*, 1990; *Chahine*, 1992; *Harries*, 1997]. *Raval and Ramanathan* [1989] defined the classic "greenhouse factor", G, as the difference between the upwelling flux emitted at the surface and the outgoing longwave radiation. To illustrate the role of water vapor and the far-infrared in the natural greenhouse effect we show G in Figure 4 in total and for the far-infrared, mid-infrared, and window channels. This figure clearly shows the dominance of the far-infrared in establishing the natural clear-sky greenhouse effect of the Earth.

$$G = S_U - OLR$$

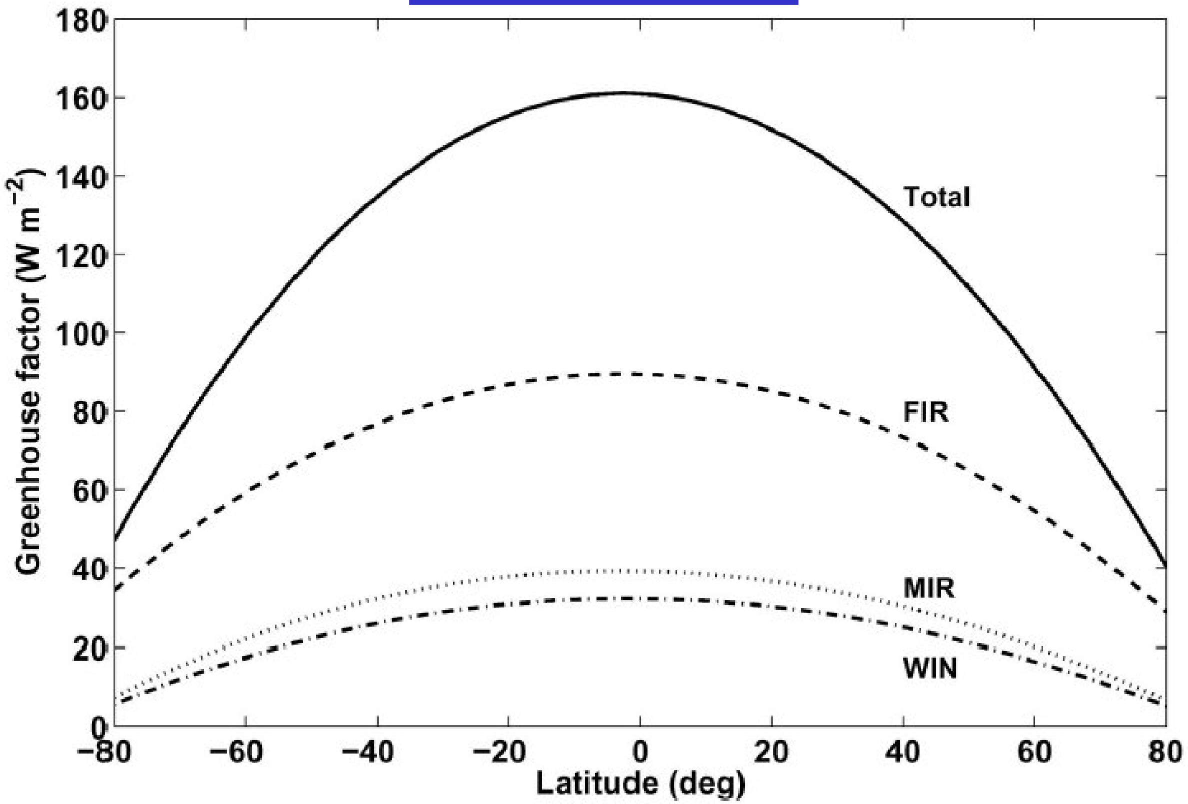
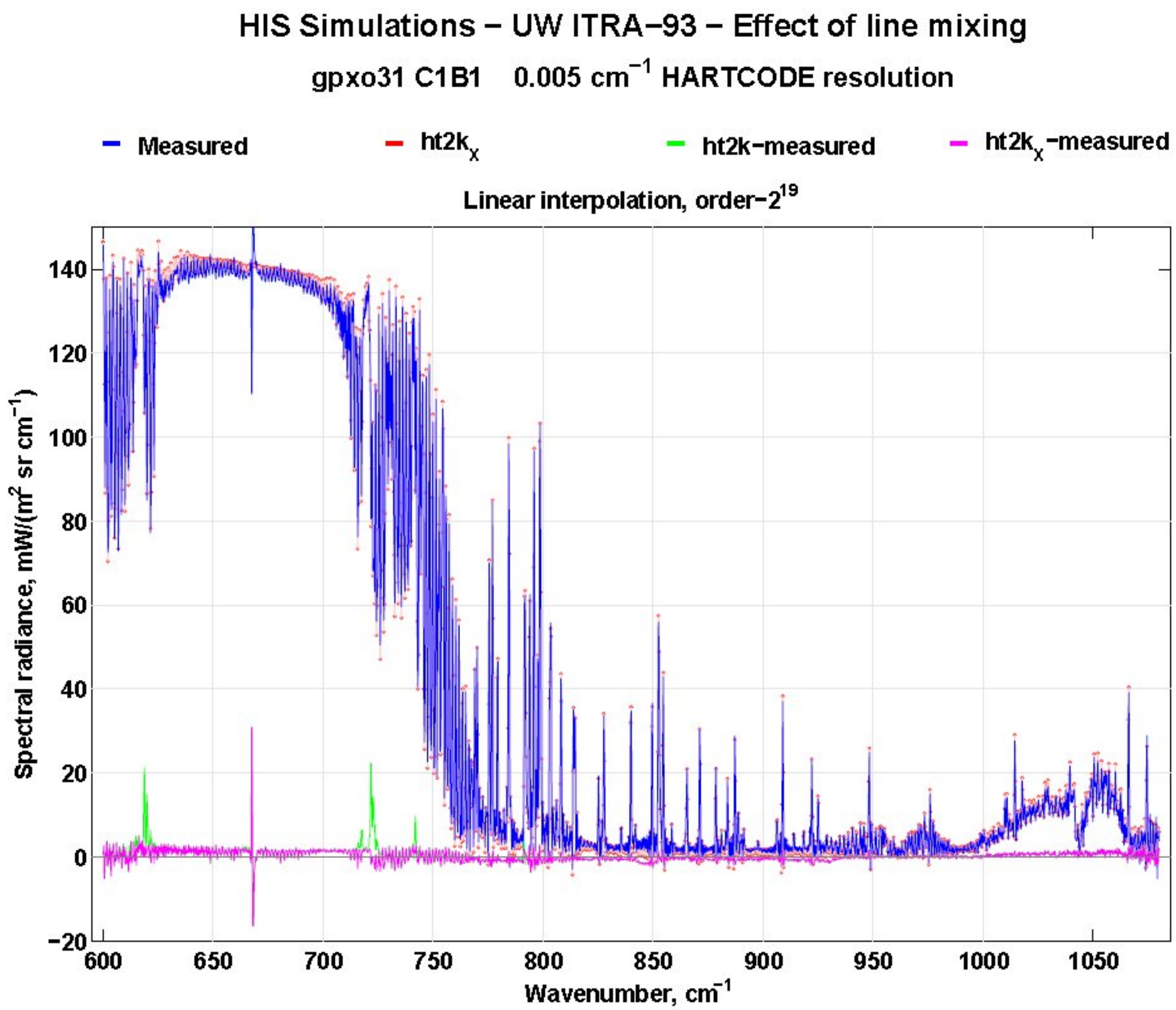
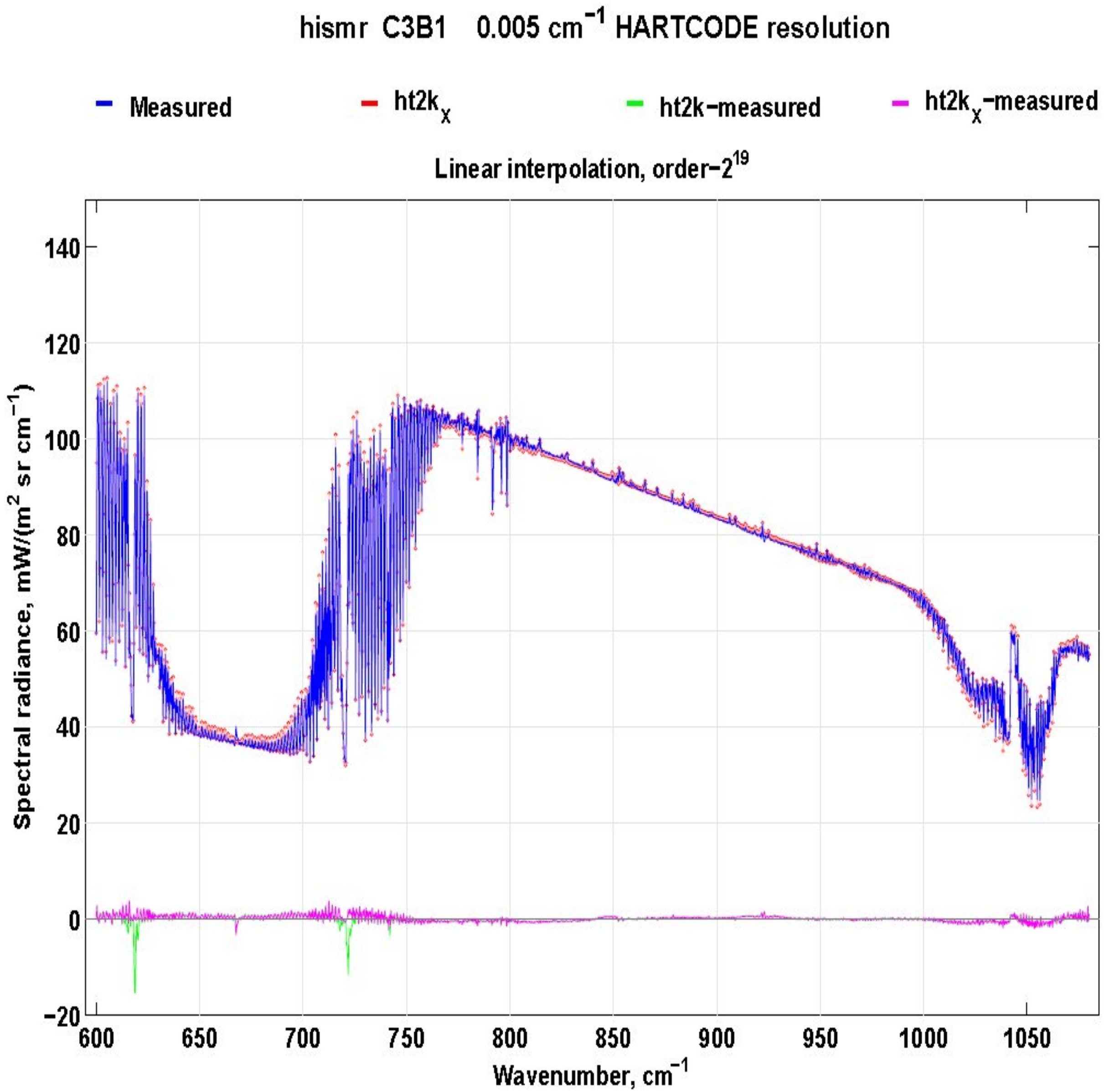


Figure 4. The classic greenhouse factor, G, for clear sky for the entire infrared and in the far-infrared, window, and mid-infrared spectral regions.

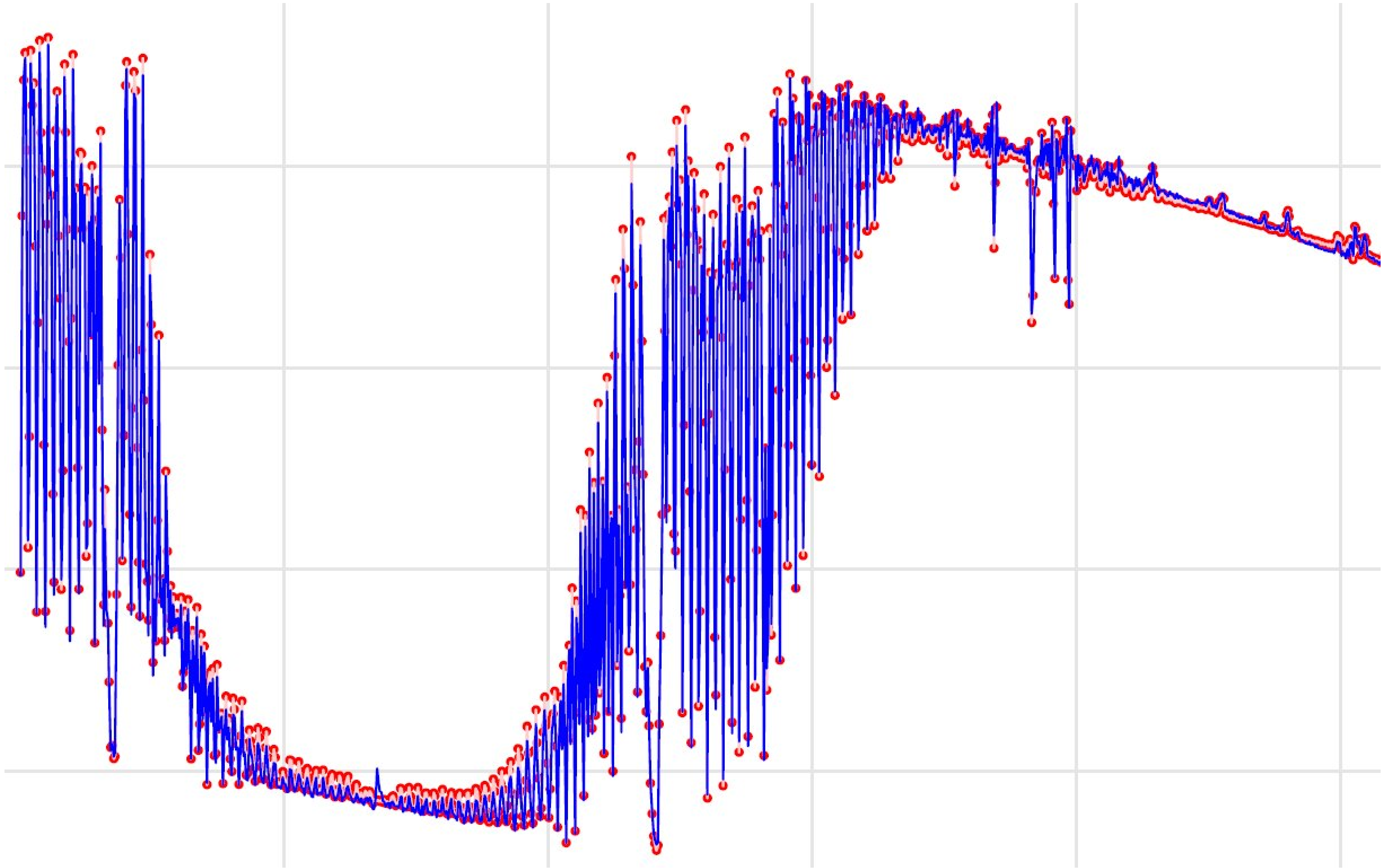
Proof of HARTCODE's ability to reproduce Fourier interferometer measurements



Looking upward from the ground



Looking downward from an airplane

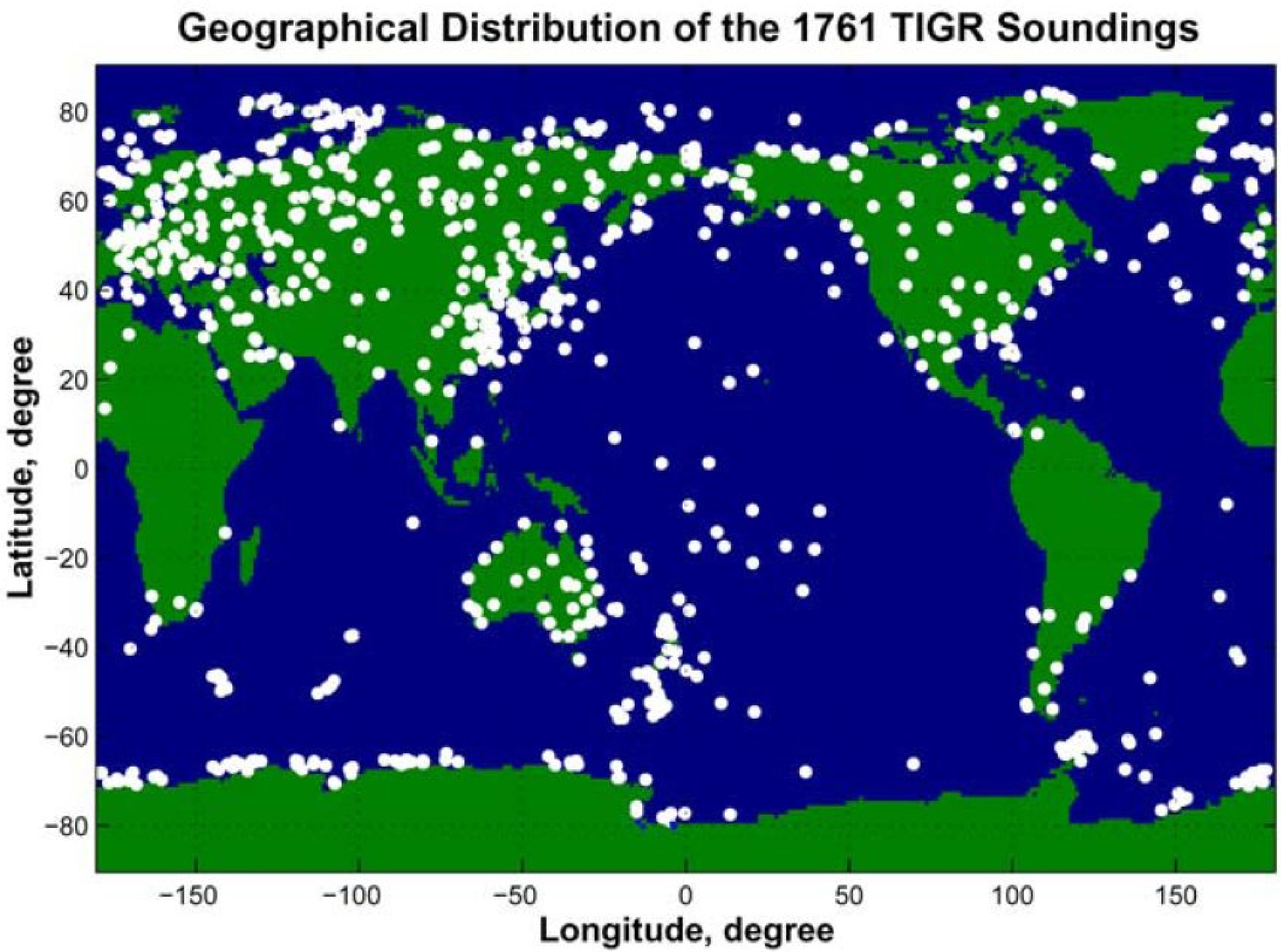


Part II. Applications of HARTCODE

The original idea was not climatology:

- To prepare a large set of IR transmittance function profiles on carefully selected model atmospheres — >
- Evaluating of remote sensing data by simple and fast interpolation methods, avoiding the complicated subsequent solutions of the IR transfer equation — >
- Producing the necessary atmospheric database for numerical weather prediction.

Global scale simulations



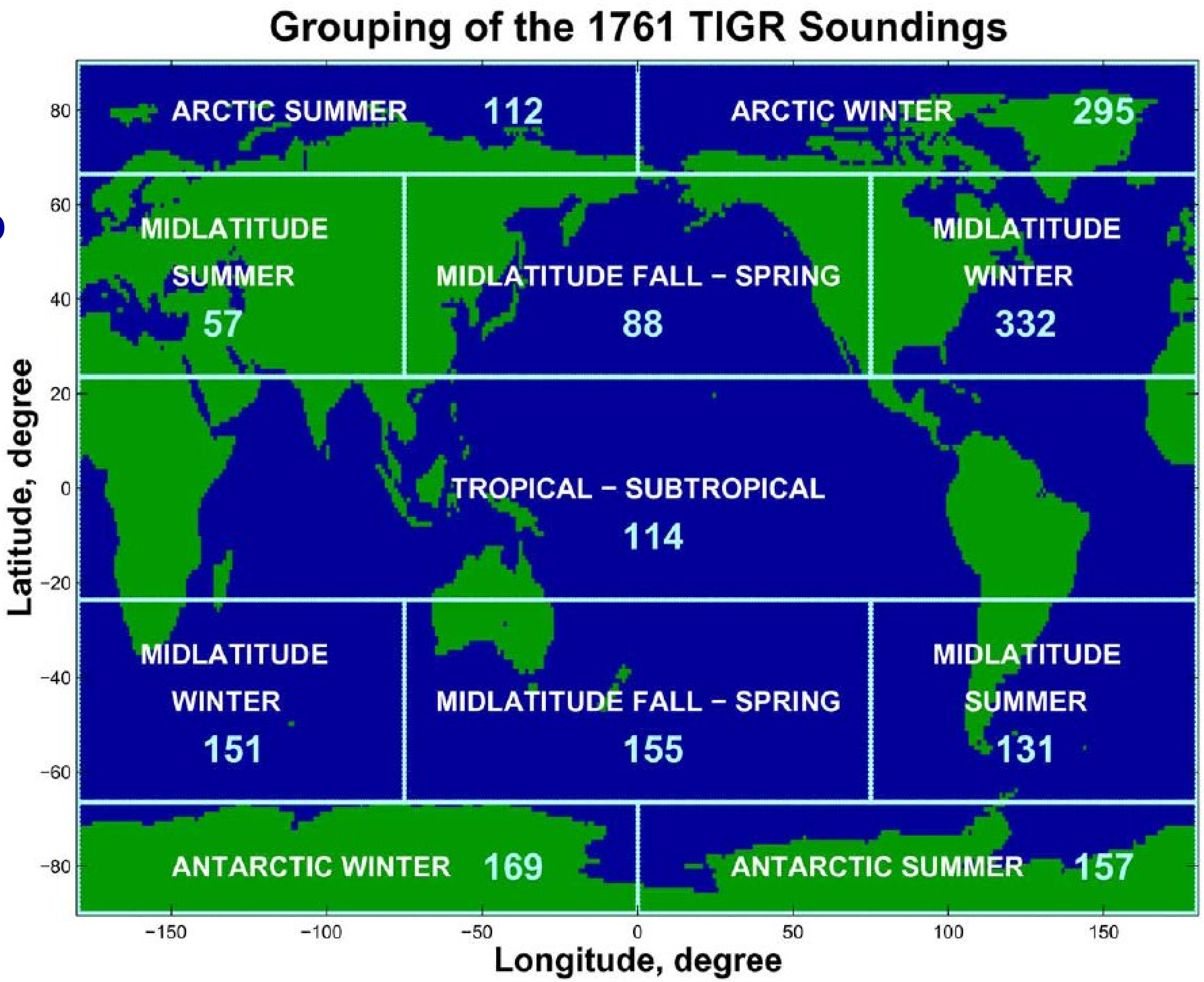
Quotes from Miskolczi (Quarterly Journal of the Hungarian Meteorological Society, QJHMS 2001):

The present study uses a subset of the TOVS Initial Guess Retrieval, (TIGR), database of 1761 radiosonde observations, Chedin and Scott, (1983). This data set contains the pressure, temperature, H_2O and O_3 mixing ratio profiles at 40 pressure levels between 0.05 and 1013 HPa. The soundings were collected over both hemispheres and over all seasons from 1976. To obtain the best retrieval on a global scale, the profiles were classified according to their geographical latitudes and the seasons. Based on the latitudinal and annual distribution, 5 latitudinal belts were selected, and in each belt, one, two or three 'seasons' were established, roughly based on the solar climate. Because of the apparent asymmetry in the global and seasonal distribution of the available solar radiation, the northern and southern hemispheres were treated separately. This classification of profiles resulted in 11 groups with a minimum of 57 profiles during the northern midlatitude summer, and a maximum of 331 profiles during the northern midlatitude winter. Further on, for practical reasons, it was necessary to reduce the number of the profiles to a reasonably small number, suitable for detailed line-by-line calculations. Due to the fact that the window channel radiances are affected mainly by the absorption of atmospheric water vapor, the selection of the individual TIGR profiles was based on the total precipitable water, *u*, effective H_2O temperature, T_e, and effective H_2O pressure, p_e. The effective values were computed by weighting the temperature and pressure profiles with the water vapor column density profile. A pre-selection based on the statistical characteristics of the 11 groups resulted in 297 profiles. The extreme profiles from each group (profiles closest to the average +/- three standard deviations) have been excluded on the basis that we did not want the regression coefficients to be affected by some statistically insignificant rare cases. After eliminating the redundancies the final set has been reduced to 228 profiles. This set contained approximately 20 profiles in each class. In Table II. the characteristics of the original data set are summarized. The selected subset of the 228 profiles has a similar statistical pattern.

Selecting 228 profiles, representing the global average

LBL SIMULATIONS USED 228 TIGR RADIOSONDE OBSERVATIONS

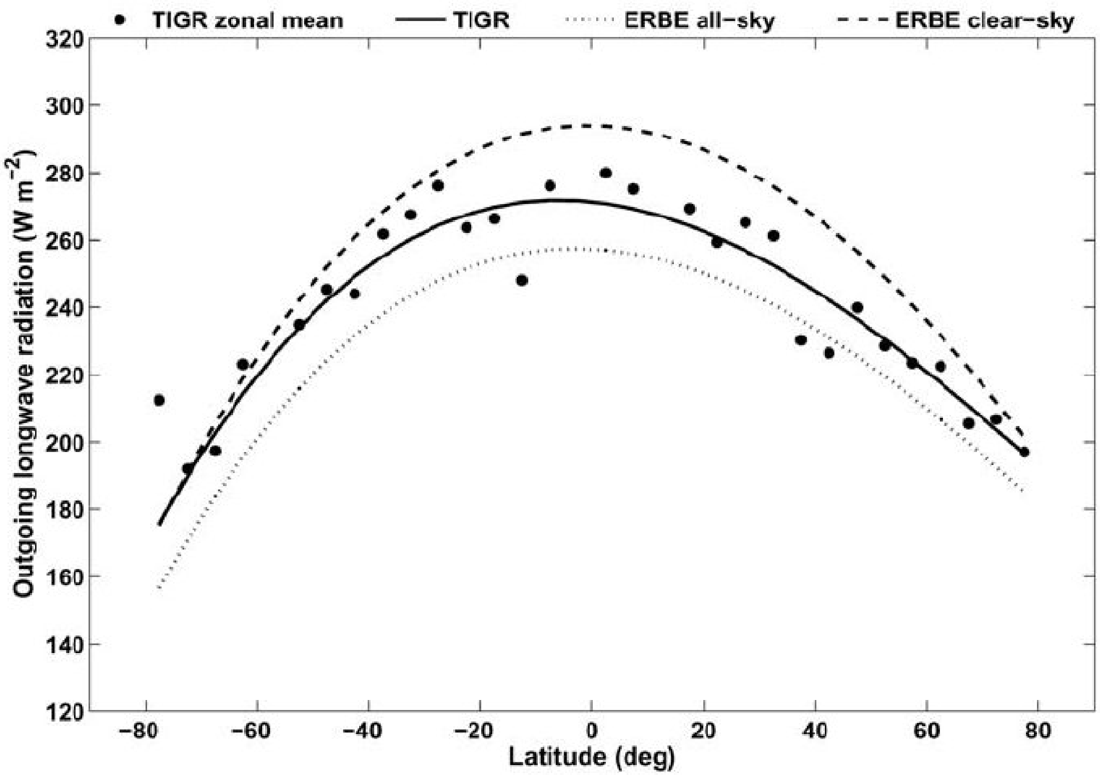
Each group contains about 20 profiles



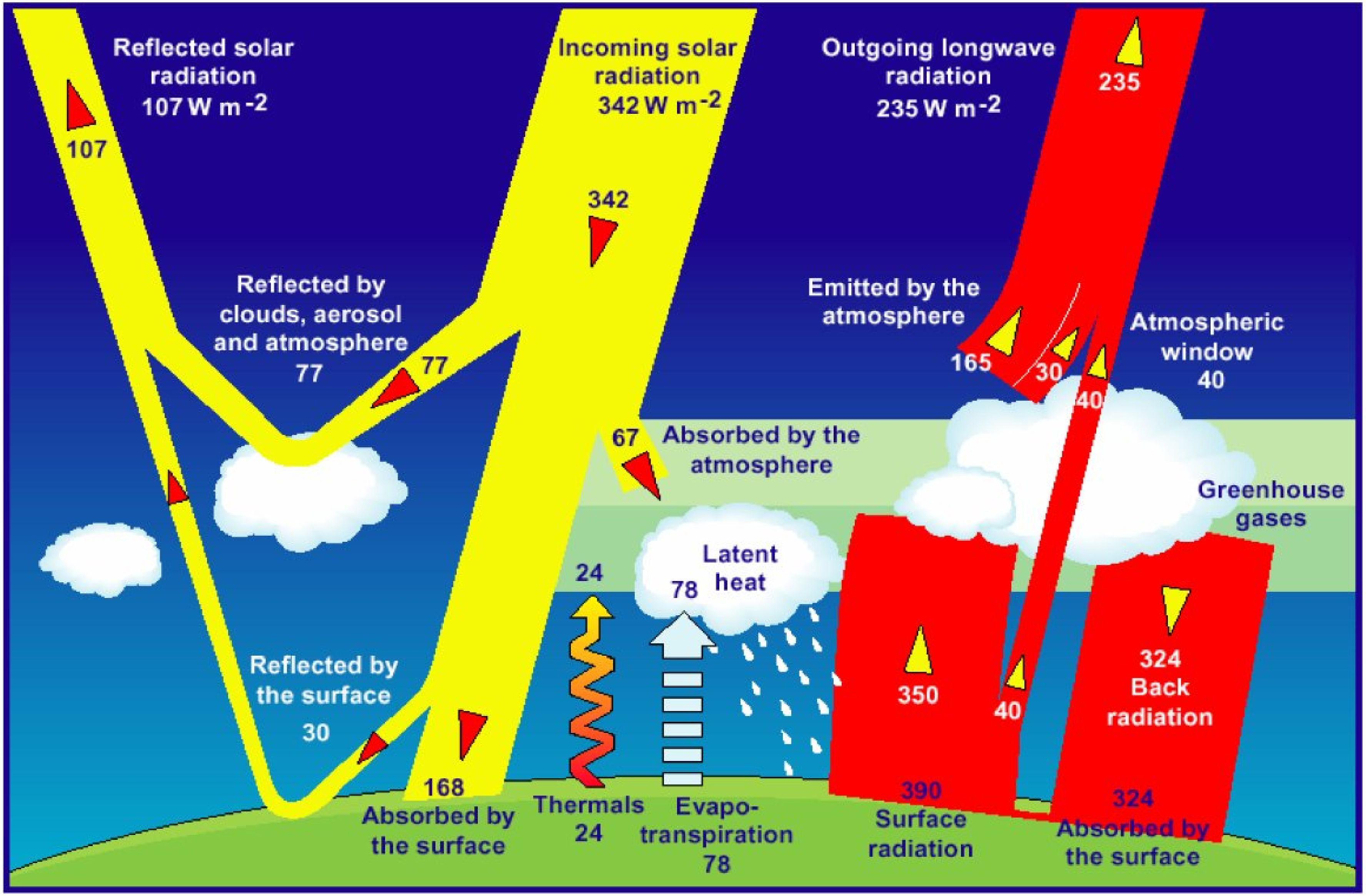
Part III. Results of HARTCODE computations on the selected 228 TIGR vertical profiles

TIGR data set are compared to the mean clear-sky and all-sky meridional distributions obtained from the ERBE measurements. The top and bottom curves in this plot were obtained by fitting a third order polynomial to the zonal averages of about 70,000 all-sky and 40,000 clear-sky annual average *OLR* measurements from the ERBS, NOAA9, and NOAA10 satellites. The markers in this figure are the TIGR OLR fluxes averaged over latitudinal belts of 5-degree widths.

OLR

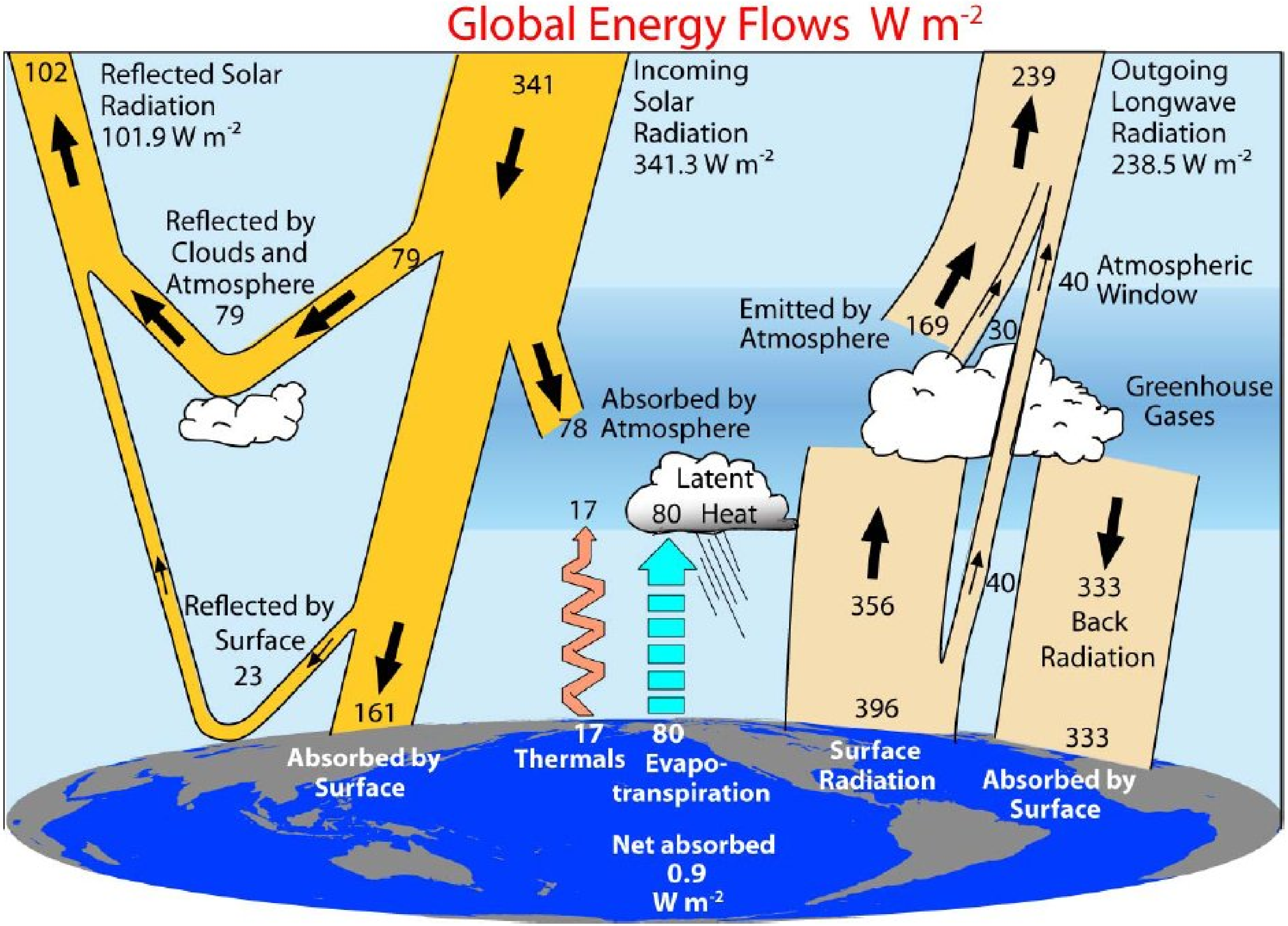


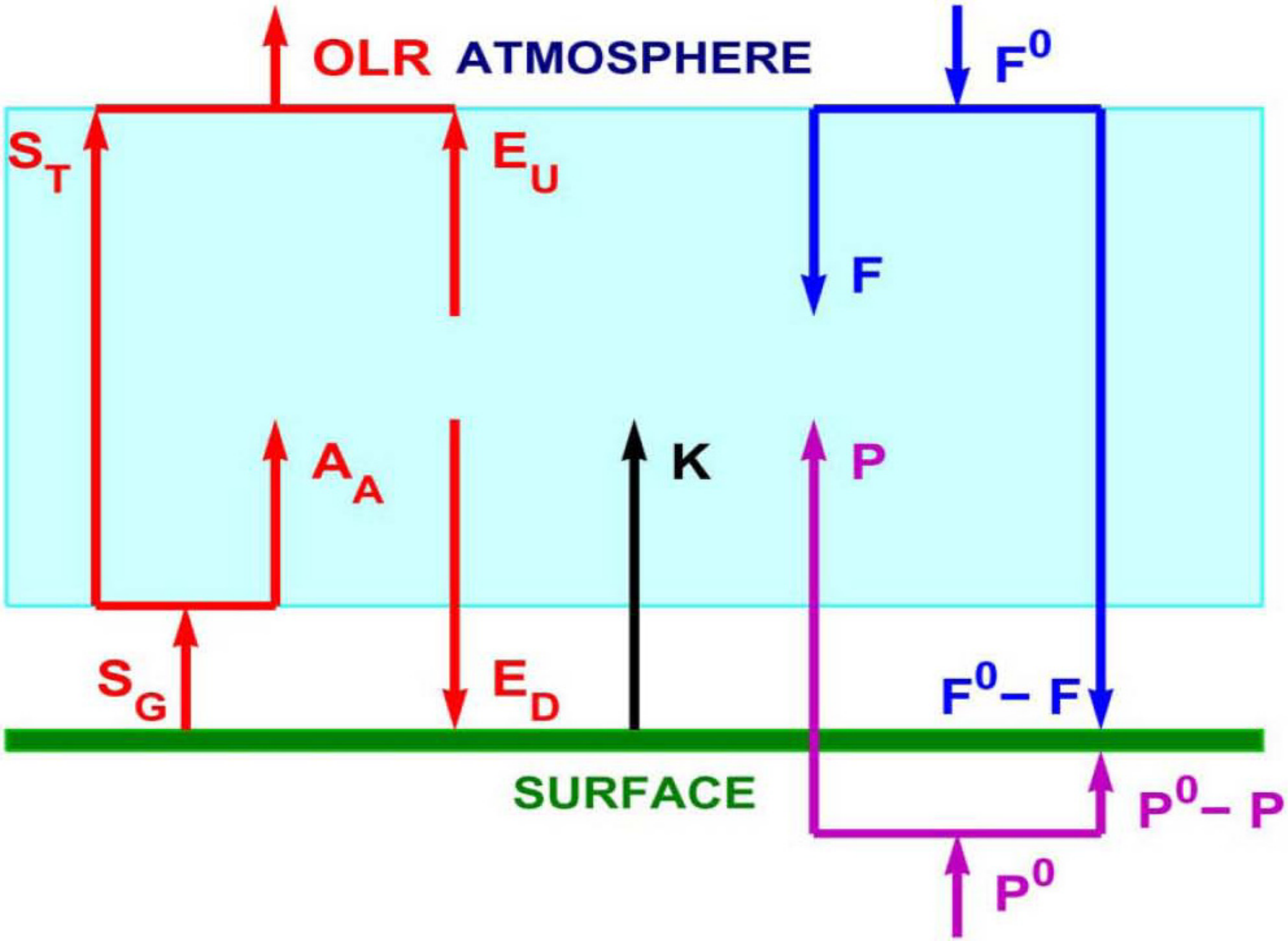
QJHMS, 2004



Source: Kiehl-Trenberth, BAMS 1997

UPDATED (Trenberth *et al.* BAMS 2009)



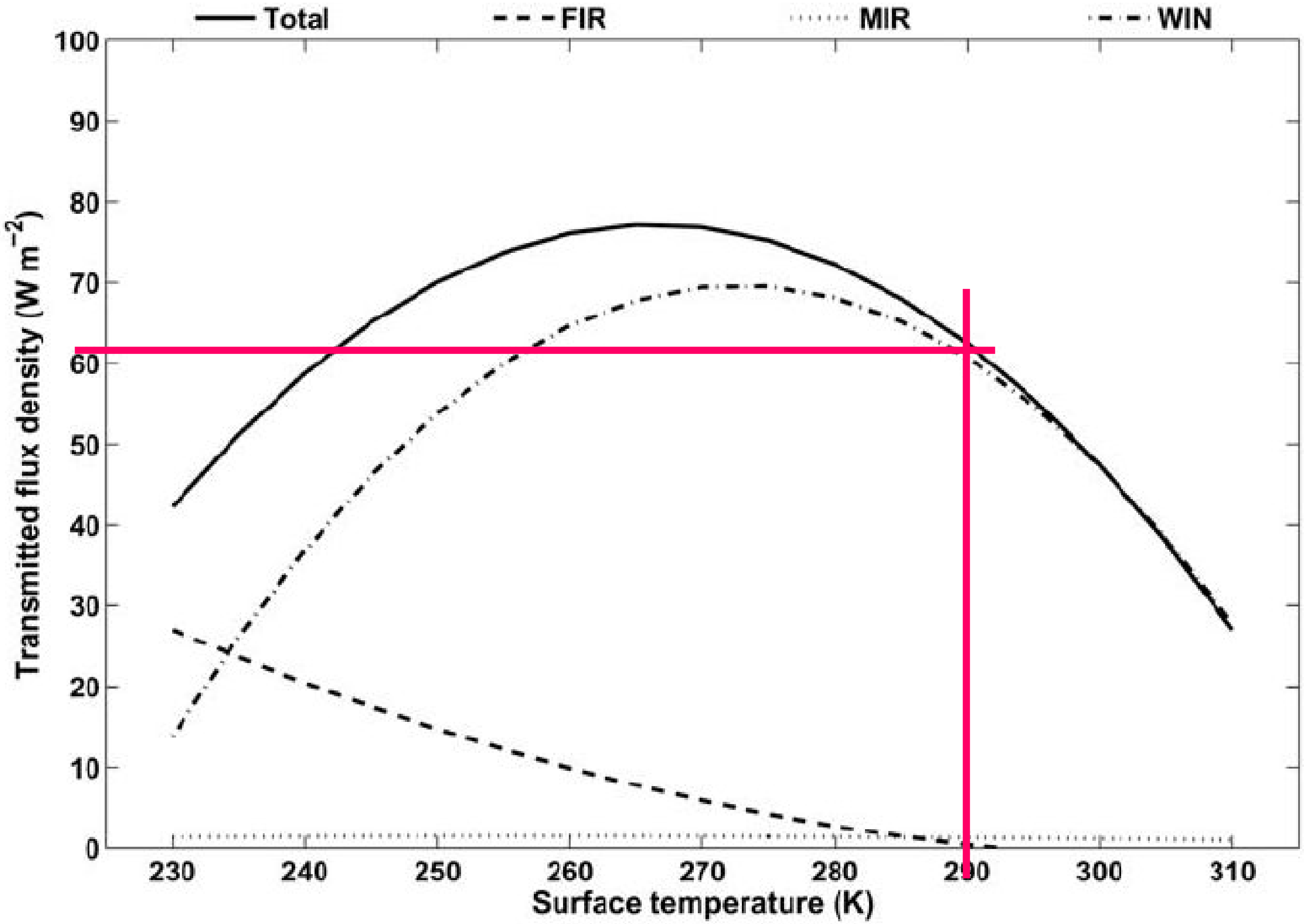


F^0=ISR*(1 – α)=OLR, α=0.30, F=atmospheric absorbed SW, K=thermals+latent, P=geothermal+ ocean/atmosphere heat exchange+industrial heat generation etc.

S_T

(T=289K)

=61 Wm^{-2}



$T_A = S_T/S_U$

(w=2.62prcm)

=0.1586

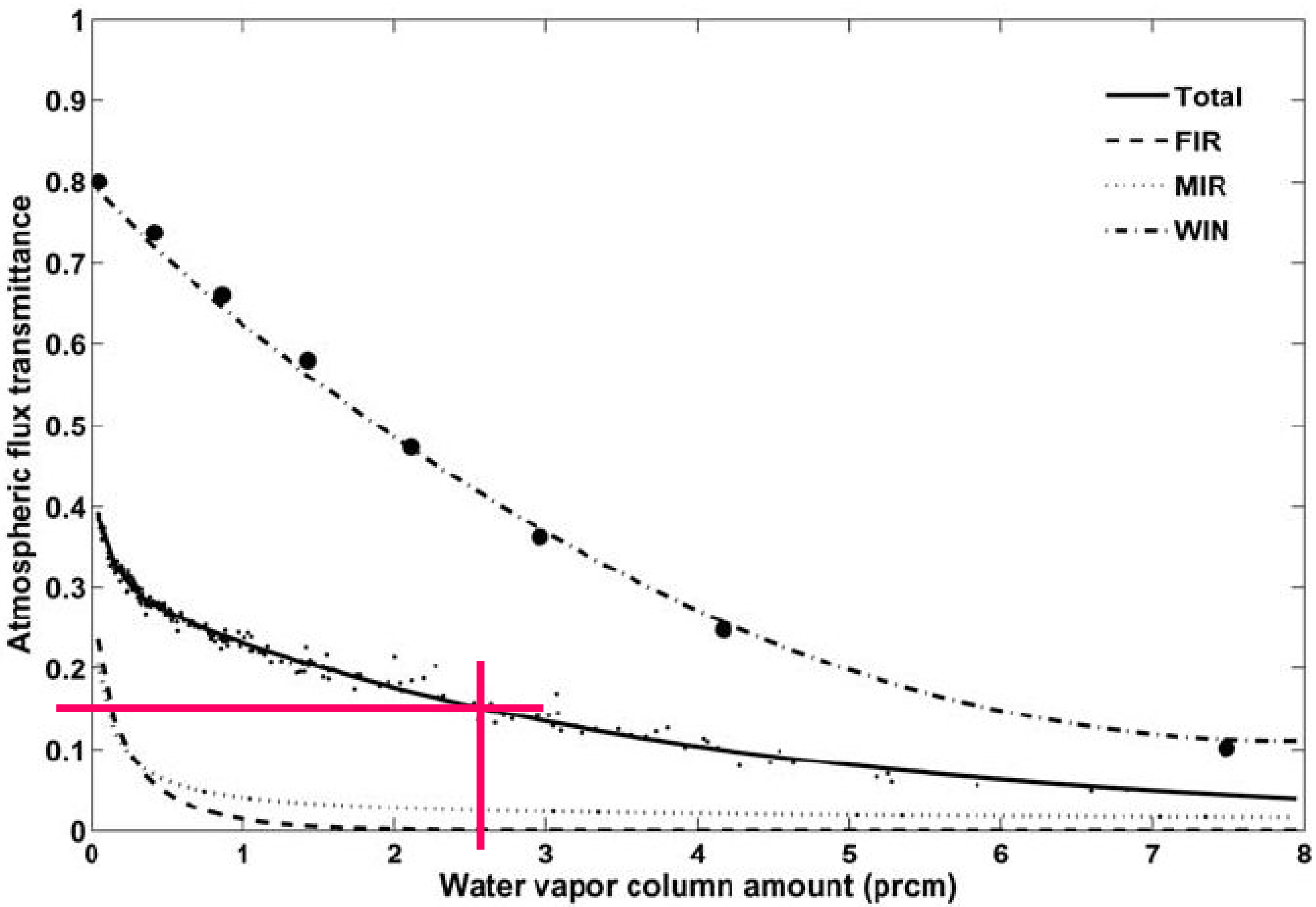


Fig. 7. Dependence of total and spectral atmospheric flux transmittance components on water vapor column amount. The large dots are the parameterized WIN Tr_A functios tested with an independent set of eight profiles.

$\tau_A = -\ln(T_A)$

$(T_A=0.1586$,

w=2.62 prcm)

= 1.868

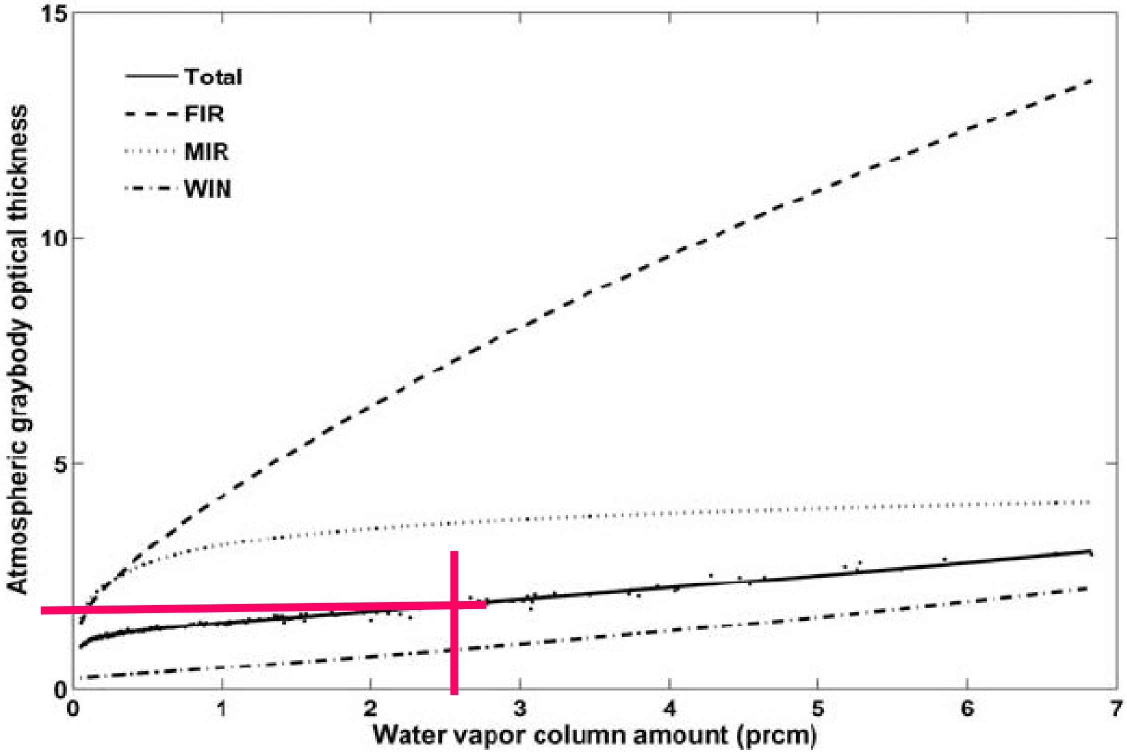


Fig. 8. Dependence of total and spectral atmospheric graybody optical thickness, $\tau_A = -\ln(S_T / S_U)$, on water vapor column amount.

By definition, the atmospheric graybody optical thickness (τ_A) is expressed as: $\tau_A = -\ln(Tr_A)$. The dependence of the total and spectral τ_A on *w* is displayed in *Fig. 8*, and the relevant parameterized formulas are given by Eqs. #5–#8. For the total τ_A the original data points are also plotted. The important message of

Two new relations:

$E_D = S_U - S_T = A_A$

$E_U = S_U/2$

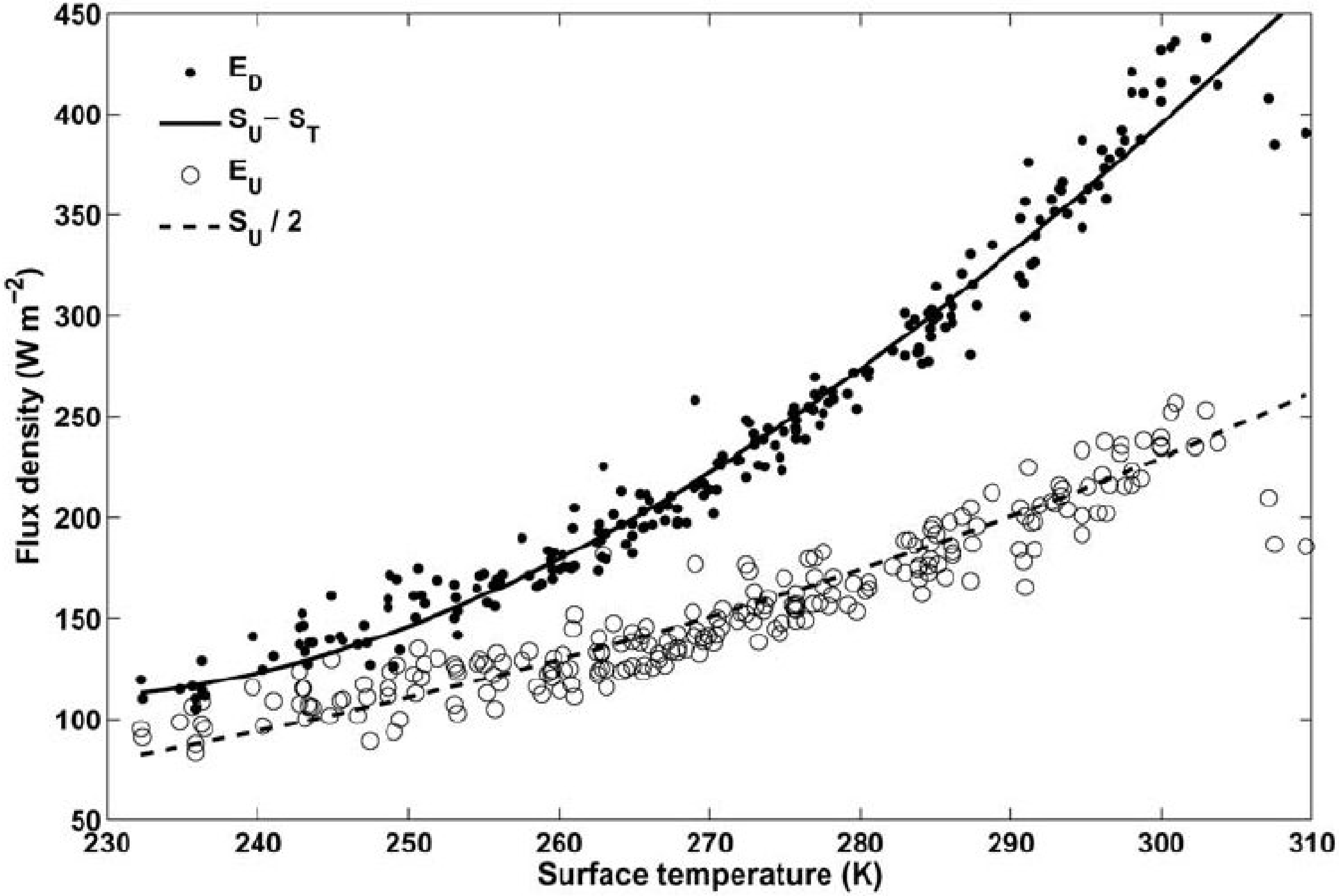
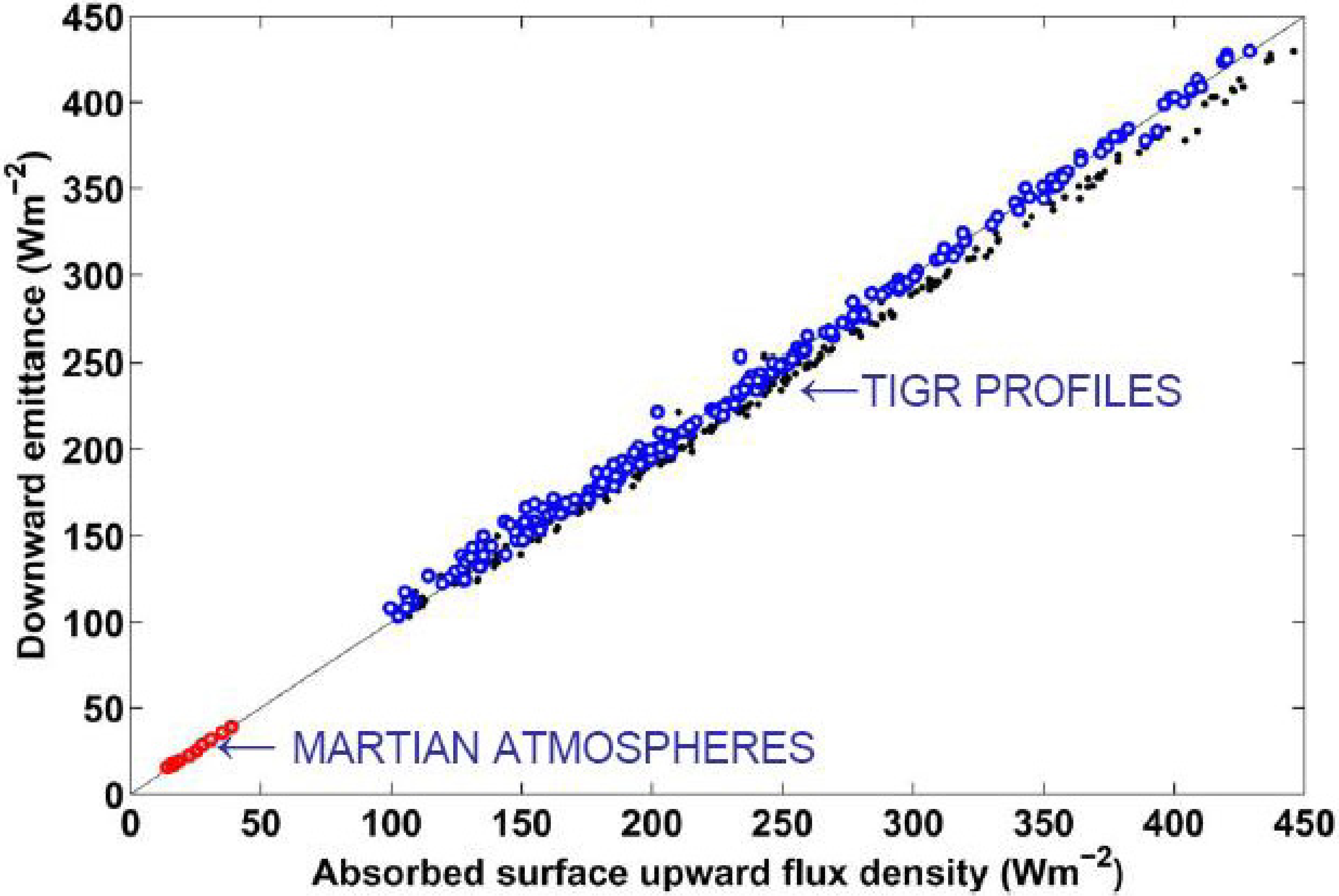


Fig. 20. Dependence of E_D and E_U fluxes on surface temperature. The solid line is a parabolic fit to the absorbed fluxes and the dashed line is the fit to the half of the surface upward fluxes.



$E_D = A_A$ independently of the thermal structure and greenhouse gas content of the atmosphere.

Summing up Miskolczi's QJHMS 2004:

Two new relations ... :

$S_U = 2E_U$, $E_D = A_A$

- In the average sense the atmosphere is very close to the radiative equilibrium, and, as a consequence, the zonal and global average upward emittance is about half of the average surface upward flux density. This fact is supported by the recent assessment of the Earth's annual global mean energy budget by *Kiehl* and *Trenberth* (1997). Their estimates of S_U and E_U are 390 and 195 W m^{-2}, respectively.

- As a consequence of the Kirchoff's law, within the clear atmosphere the downward emittance is approximately equal to the absorbed flux density. Based on our data set, the global average clear-sky downward atmospheric emittance is 311.4 W m^{-2}, while the global average of the absorbed radiation by the clear-sky is 311.9 W m^{-2}. This equivalence – for the highly variable atmospheric emission spectra and for global scale – was not shown

...and the Earth's global average IR optical depth (τ_A) :

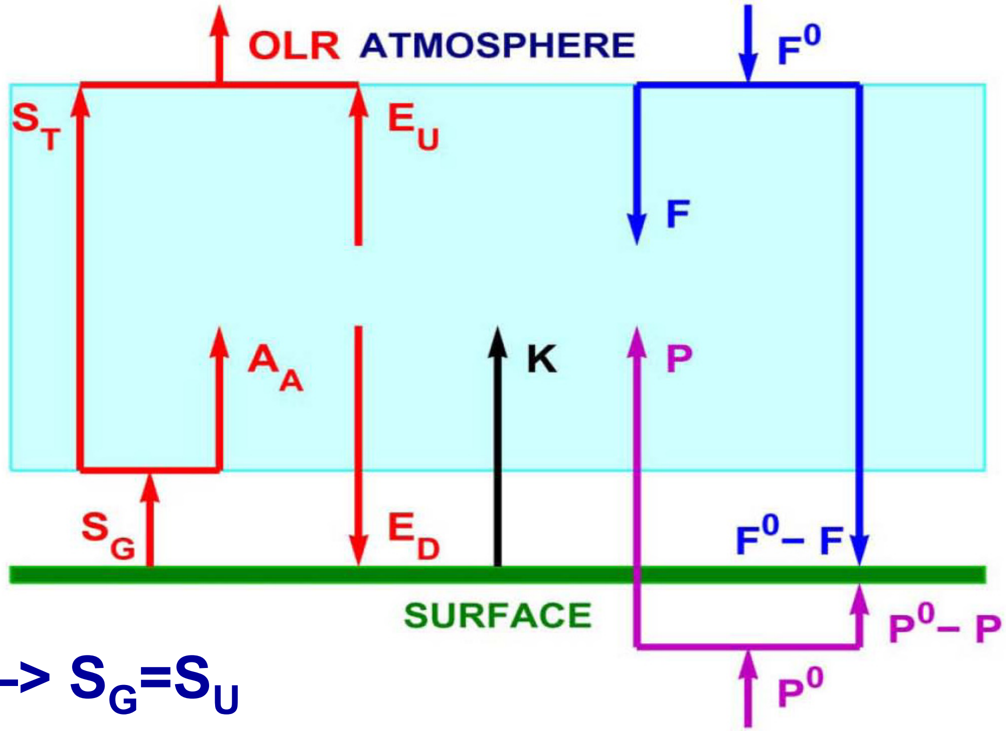
$$\bar{T}_A(\Delta\nu,\mu) = \frac{1}{\Delta\nu}\int_{\Delta\nu} \exp\left[-\sum_{l=1}^{L}\sum_{i=1}^{N}\left[c^{i,l} + k_\nu^{i,l}\right]\frac{u^{i,l}}{\mu^l}\right]d\nu$$

$$\tilde{T}_A(\Delta\nu) = \int_{2\pi} \bar{T}_A(\Delta\nu,\mu)d\varpi$$

$$T_A = \frac{1}{\sigma t_A^4}\sum_{j=1}^{M}\pi B(\Delta\nu_j, t_A)\tilde{T}_A(\Delta\nu_j)$$ —> $\tau_A = -\ln(T_A)$

$$\tau_A = 1.868$$

Part IV. Consequences (QJHMS 2007)



$\varepsilon_G = 1 \longrightarrow S_G = S_U$

NET ATMOSPHER (1) $F + P + K + A_A - E_D - E_U = 0$

NET SURFACE (2) $F^0 + P^0 + E_D - F - P - K - S_U = 0$

(3) $F^0 + P^0 = OLR$

(4) $A_A = E_D$.

Inference on

(1) $F + P + K + \boxed{A_A - E_D} - E_U = 0$ and

(2) $F^0 + P^0 - S_U + E_D \boxed{- F - P - K} = 0$:

- Eq (1) (atmosphere) becomes:

  (5) $E_U = F + K + P$ ("1a")

- Eq (2) (surface) turns to:

  (6) $S_U - OLR = E_D - E_U$ ("2a")

(6) $G = S_U - OLR = E_D - E_U$

This equation describes the equality of a net upward and a net downward flux density.

S_U – OLR heats the atmosphere, $E_D - E_U$ is the answer of the atmosphere to this drive: it maintains the energetic equilibrium at the surface.

The presence of these two longwave flux densities in the air is the consequence of the atmospheric IR-active gases, GHGs.

The source of $(S_U - OLR) + (E_D - E_U)$ is the incoming available $F^0 + P^0$ flux.

Therefore we can write (energy conservation):

(7) $(S_U - OLR) + (E_D - E_U) = F^0 + P^0 = OLR$

This equation assumes the *optimal (that is, maximal*) conversion of F^0 into OLR.

Using (6), from (7) we get:

(8) $S_U = 3\ OLR/2$

—> $G = S_U - OLR = E_D - E_U = S_U/3$

$g = G/S_U = 1/3$.

FRONTIERS OF CLIMATE MODELING

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5

The radiative forcing due to clouds and water vapor

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5.4 Atmospheric greenhouse effect: global and regional averages

Consider first a one-dimensional system with the surface emitting like a black body. The clear-sky outgoing longwave radiation (F_c) and G_a are related by Equation (5.3), where T_s is the surface temperature, and G_a, by definition is given by

$$F_c = \sigma T_s^4 - G_a \quad (\, OLR = S_U - G \,) \tag{5.3}$$

The global average G_a is 131 W m^{-2} or the normalized g_a is 0.33, i.e., the atmosphere reduces the energy escaping to space by 131 W m^{-2} (or by a factor of 1/3). The ocean regions have a slightly larger greenhouse effect (0.35 for ocean

g =1/3 empirical fact (2006)

First conclusion

- **The $g = G/S_U$ clear-sky normalized greenhouse factor of the Earth is not coincidently, but necessarily equals to 0.333 ;**
- **its critical (or equilibrium) value is 1/3 .**
- **This is a direct arithmetic consequence of Miskolczi's $A_A = E_D$ equation.**

Conditions of $A_A = E_D$:

- Thermal equilibrium at the lower boundary (surface): $t_S = t_A$,
- Local Thermodynamic Equilibrium (LTE) in the atmosphere ,
- Exact solution of the radiation transfer equation with correct boundary conditions.

Part V. Historical perspective

876 Mr. E. A. Milne *on Radiative Equilibrium:* **(1922)**

outer limit of the atmosphere; if ρ is the density at height h, $k(h)$ the mass-absorption coefficient, then

$$\tau(h) = \int_h^\infty k(h)\rho\, dh.$$

Let $I(\tau)$ be the intensity of radiation at τ in a direction θ with the outward vertical, where $0 \leq \theta \leq \frac{1}{2}\pi$; and let $I'(\tau)$ be the intensity at ψ with the inward vertical, where $0 \leq \psi \leq \frac{1}{2}\pi$. Assume the material is *grey* (*i.e.* has an absorption coefficient the same for all the wave-lengths that are important—in this case the wave-lengths that are predominant in the low-temperature radiation considered). Let $B(\tau)$ be the intensity of black body radiation for the temperature ruling at the point τ; and let $\pi F(\tau)$ be the net upward flux of energy per unit area across a horizontal plane at τ. Then

$$\cos\theta \frac{dI}{d\tau} = I - B, \quad . \quad . \quad . \quad . \quad . \quad . \quad (1)$$

$$\cos\psi \frac{dI'}{d\tau} = B - I', \quad . \quad . \quad . \quad . \quad . \quad . \quad (2)$$

$$\frac{dI}{dt} = I - B, \quad \frac{dI'}{dt} = B - I', \quad . \quad . \quad . \quad (5), (6)$$

$$F(t) = I(t) - I'(t); \quad . \quad . \quad . \quad . \quad . \quad . \quad (7)$$

and the equation for t_2 is

$$I(t_2) - I'(t_2) - I(0) = 0, \quad . \quad . \quad . \quad . \quad . \quad . \quad (8)$$

since the incident radiation $I'(0)$ is zero †. Solving (5) and (6) with the assumption that the air near the ground has the same temperature as the ground and that the earth radiates like a black body, we find

$$I(t) = e^{t}\int_{t}^{t_1} B(t)e^{-t}dt + B(t_1)e^{-(t_1-t)}dt, \quad . \quad . \quad (9)$$

$$I'(t) = e^{-t}\int_{0}^{t} B(t)\, e^{t}dt. \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (10)$$

§ 4. *The problem in principle.*—As a contribution towards this, it is proposed in this paper to consider the theory of the radiative equilibrium of a mass of absorbing and radiating material subject to insolation. The material is supposed to be stratified in parallel planes, and to be subject at its outer boundary to a parallel beam of incident radiation. The latter will be supposed in the first instance to be normal to the surface; later we shall examine the effect of oblique incidence. The material will be taken in the first instance to be grey; but later we shall suppose that there may be one coefficient of absorption for the incident radiation, another coefficient for the low-temperature radiation emitted by the material itself. Further, we shall assume the material to be infinitely thick, and to be in radiative equilibrium throughout its mass. The assumption of infinite thickness involves little or no loss of generality; we could, if we liked, consider a mass of finite thickness with an inner boundary consisting of a black radiating surface, but since our results will only involve the optical thickness, we need only suppose the absorption coefficient or the density to become suddenly very large at an assigned depth in order to deduce the case of an inner boundary from the solution for an infinitely thick slab of material.

The material being in a steady state must emit energy at its outer boundary equal to the incident radiation. Across any plane parallel to the surface there will be a net outward flux of radiation derived from the material just balancing the inward flux of the residual solar radiation. In the far interior the latter will be greatly attenuated, and consequently the outward flux there must be small too. We should expect, therefore, that the temperature gradient in the far interior would be small; and this proves to be the case. In fact, *not only is there a definite limiting temperature at the outer boundary, as in the Schwarzschild case, but there is also a definite limiting temperature in the far interior.* This is one of the most interesting characteristics of the model we are discussing.

$$H = \int_0^{\infty} \int_{4\pi} I_\nu \cos(\vartheta) d\omega\, d\nu$$ Eddington flux

$$dB(\bar{\tau}) / d\bar{\tau} = 3H / (4\pi)$$ Radiative equilibrium

$$B(\bar{\tau}) = 3H\, \bar{\tau} / (4\pi) + B_0$$ Planck blackbody source function

$$H = \pi \left[\bar{I}^+(\bar{\tau}) - \bar{I}^-(\bar{\tau}) \right]$$ Planparalel hemispheric S-M. eq.

AT THE TOP $\bar{I}^-(\bar{\tau}) = 0$, $\bar{\tau} = 0$ AND AT THE BOTTOM $\bar{\tau} = \infty$:

$$\frac{H}{\pi} = \frac{3}{2} \int_0^{\infty} \frac{3H}{4\pi} \bar{\tau}' e^{-\frac{3}{2}\bar{\tau}'} d\bar{\tau}' + \frac{3}{2} \int_0^{\infty} B_0 e^{-\frac{3}{2}\bar{\tau}'} d\bar{\tau}' = \frac{H}{2\pi} + B_0$$

$$\boxed{B_0 = \frac{H}{2\pi}}$$

$$S_A = OLR(1 + \tilde{\tau}_A)/2, \qquad t_A^4 = t_E^4(1 + \tilde{\tau}_A)/2$$

$$S_G = OLR(2 + \tilde{\tau}_A)/2, \qquad t_G^4 = t_E^4(2 + \tilde{\tau}_A)/2$$

- *It can be realized that the classical semi-infinite Eddington solution is not valid to the Earth's atmosphere.*
- *The equations can be solved with real finite boundary conditions.*
- *The gift for this efforts was an analytic formula for the $f(\tau)$ transfer function.*

$$\bar{I}^+(0) = B_G e^{-\frac{3}{2}\bar{\tau}_A} + \frac{3}{2}\int_0^{\bar{\tau}_A} B(\bar{\tau}')e^{-\frac{3}{2}\bar{\tau}'}d\bar{\tau}', \tag{B1}$$

and

$$\bar{I}^-(0) = 0. \tag{B2}$$

Putting Eq. (B1) and Eq. (B2) into the $H(\bar{\tau}) = \pi(\bar{I}^+ - \bar{I}^-)$ equation, and substituting the source function with $B(\bar{\tau}) = 3H\bar{\tau}/(4\pi) + B_0$ in the upward hemispheric mean radiance we get:

$$\frac{H}{\pi} = B_G e^{-\frac{3}{2}\bar{\tau}_A} + \frac{3}{2}\int_0^{\bar{\tau}_A} \frac{3H}{4\pi}\bar{\tau}'e^{-\frac{3}{2}\bar{\tau}'}d\bar{\tau}' + \frac{3}{2}\int_0^{\bar{\tau}_A} B_0 e^{-\frac{3}{2}\bar{\tau}'}d\bar{\tau}'. \tag{B3}$$

The two definite integrals in the second and third terms of the right hand side of Eq. (B3) must be evaluated:

$$\frac{3}{2}\int_{0}^{\overline{\tau}_A}\frac{3H}{4\pi}\overline{\tau}' e^{-\frac{3}{2}\overline{\tau}'}d\overline{\tau}' = -\frac{H}{4\pi}(2e^{-\frac{3}{2}\overline{\tau}_A} - 2 + 3\overline{\tau}_A e^{-\frac{3}{2}\overline{\tau}_A}), \tag{B4}$$

$$\frac{3}{2}\int_{0}^{\overline{\tau}_A} B_0 e^{-\frac{3}{2}\overline{\tau}'}d\overline{\tau}' = B_0(1 - e^{-\frac{3}{2}\overline{\tau}_A}). \tag{B5}$$

After putting back Eqs. (B4) and (B5) into Eq. (B3) we get:

$$\frac{H}{\pi} = B_G e^{-\frac{3}{2}\overline{\tau}_A} - \frac{H}{4\pi}(2e^{-\frac{3}{2}\overline{\tau}_A} - 2 + 3\overline{\tau}_A e^{-\frac{3}{2}\overline{\tau}_A}) + B_0(1 - e^{-\frac{3}{2}\overline{\tau}_A}). \tag{B6}$$

Rearranging Eq. (B6) and using the $\tilde{\tau}_A = (3/2)\bar{\tau}_A$ notation for the total flux optical depth, πB_0 can be expressed as:

$$\pi B_0 = \frac{\frac{H}{2}\left[1 + \tilde{\tau}_A e^{-\tilde{\tau}_A} + e^{-\tilde{\tau}_A}\right] - \pi B_G e^{-\tilde{\tau}_A}}{1 - e^{-\tilde{\tau}_A}} . \qquad \text{(B7)}$$

This B_0 in the $B(\bar{\tau}) = 3H\bar{\tau}/(4\pi) + B_0$ equation will give the general form of the source function profile:

$$\pi B(\tilde{\tau}) = \frac{\frac{H}{2}\left[1 + \tilde{\tau} + (\tilde{\tau}_A - \tilde{\tau} + 1)e^{-\tilde{\tau}_A}\right] - \pi B_G e^{-\tilde{\tau}_A}}{1 - e^{-\tilde{\tau}_A}} . \qquad \text{(B8)}$$

$$\pi B_G = OLR \frac{1 + \tilde{\tau}_A + e^{-\tilde{\tau}_A}}{2} = \frac{OLR}{f} .$$

General solution (MF QJHMS, 2007):

RADIATIVE EQUILIBRIUM IN FINITE MEDIUM

$dB(\bar{\tau})/d\bar{\tau} = 3H/(4\pi)$ $\quad B(\bar{\tau}) = 3H\,\bar{\tau}/(4\pi) + B_0$ $\quad H = \pi\left[\bar{I}^+(\bar{\tau}) - \bar{I}^-(\bar{\tau})\right]$

Boundary conditions

$\tilde{\tau}_A$ **and** $S_G = \pi B_G$

$$\pi B_0 = \frac{\frac{H}{2}\left[1 + \tilde{\tau}_A e^{-\tilde{\tau}_A} + e^{-\tilde{\tau}_A}\right] - \pi B_G e^{-\tilde{\tau}_A}}{1 - e^{-\tilde{\tau}_A}}$$

Transfer function

$f = 2/(1 + \tau_A + T_A)$

$$\pi B(\tilde{\tau}) = \frac{H}{2A}\left[\frac{2}{f} - (\tilde{\tau}_A - \tilde{\tau})A\right] - \frac{\pi B_G T_A}{A}$$

Energy minimum principle

$\pi dB_0(\tilde{\tau}_A)/d\tilde{\tau}_A = 0$

$$S_U = \frac{OLR}{f} \rightarrow t_G^4 = \frac{t_E^4}{2}(1 + \tau_A + T_A)$$

The surface temperature depends also on the flux transmittance T_A.

In Eq. (8) we got:

$$S_U = (3/2)OLR$$

Now we have:

$$(9) \quad S_U = (3/2)OLR = OLR / f ,$$

with $(10) \quad f = 2/(1 + \tau_A + \exp(-\tau_A))$.

Taking together

$S_U = 3\ OLR / 2$ and $S_U = 2\ E_U$,

we get

(11) $$f = OLR/S_U = 3/5 + 2T_A/5 ,$$

that is,

$$2/(1 + \tau_A + \exp(-\tau_A)) = 3/5 + 2\exp(-\tau_A)/5 ,$$

which gives for τ_A as general solution:

$$\tau_A = 1.867561 \ldots$$

Observation

(HARTCODE computation on TIGR, 2004): $\tau_A = 1.868$.

Theoretical derivation

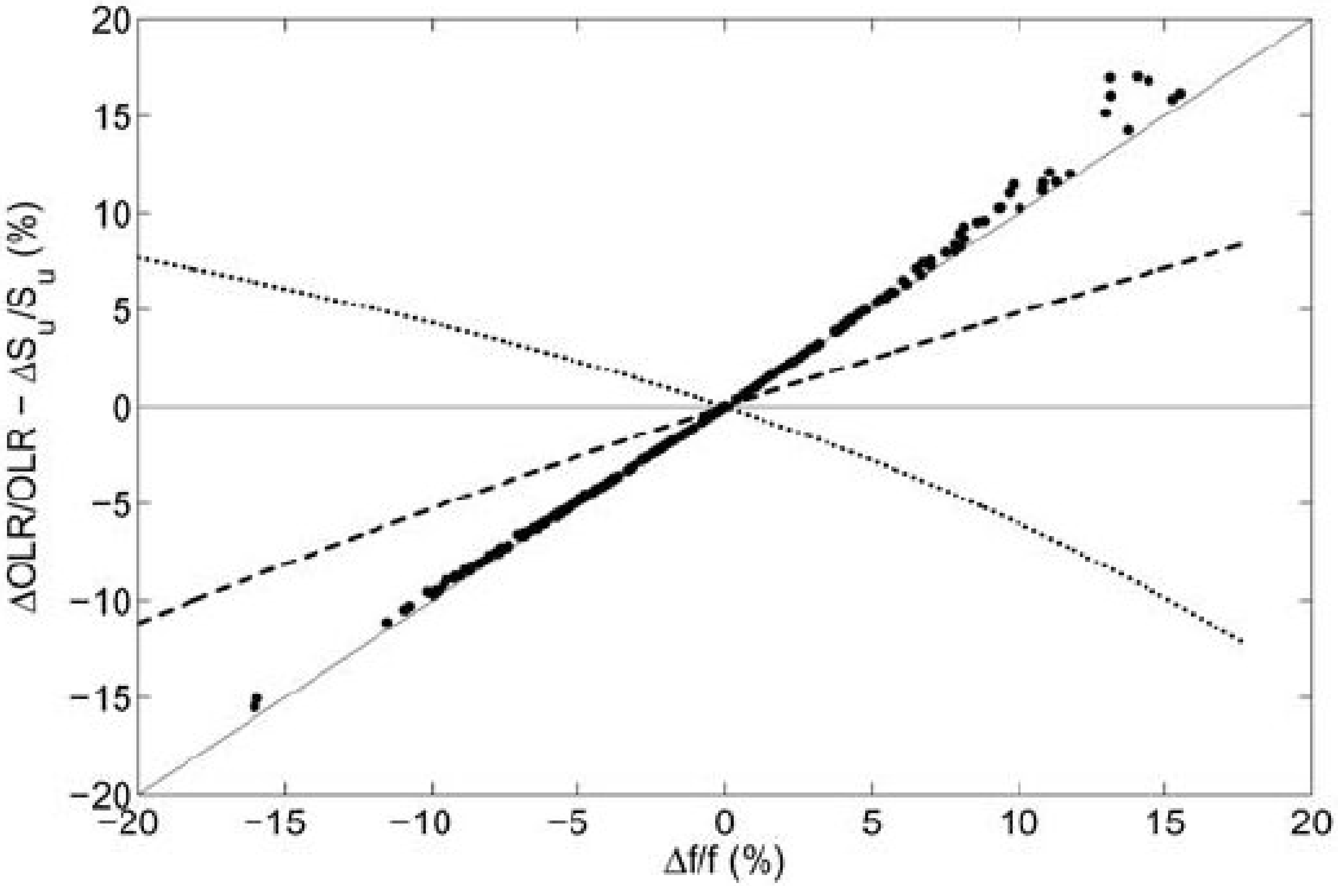
(Eqs 8-28, 2007): $\tau_A = 1.86756$.

- **The difference is less then 0.1 %.**
- **I regard this deduction of F. Miskolczi one of the most beautiful results in the history of theoretical physics.**

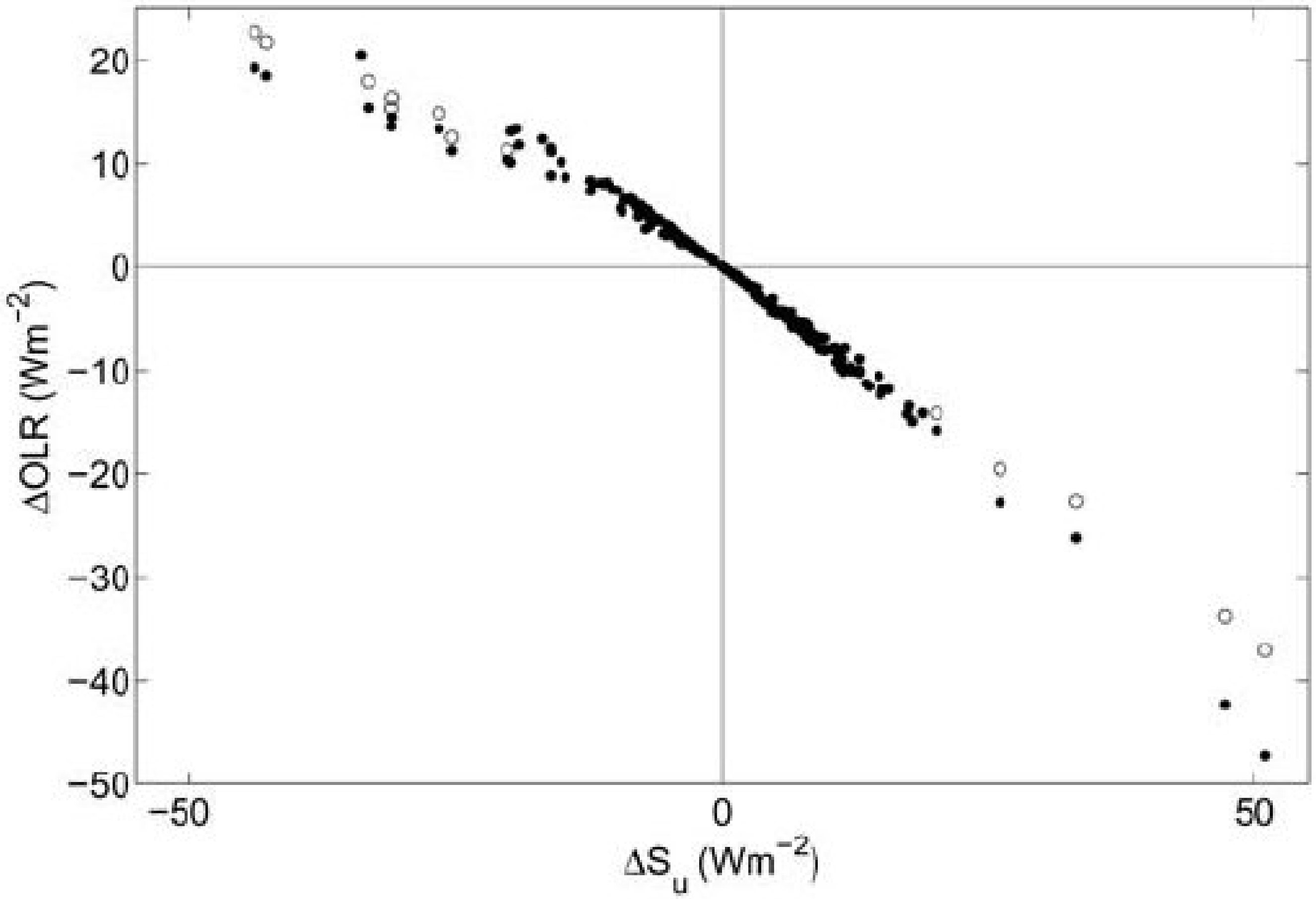
Part VI. Interpretation

- **The S_U surface upward radiation and surface temperature is connected unequivocally to the available F^0+P^0 energy source.**
- **The value of the $G=S_U–OLR$ greenhouse factor is fixed, and equals to $S_U/3$. Excess or deficit in G violates energetic constraints.**
- **The Earth-atmosphere system maintains an equilibrium greenhouse effect, with $g = G/S_U = 1/3$ normalized greenhouse factor.**

In S_U=OLR/f , any relative deviation from their equilibrium values…



… must be compensated by the system's energetic constraints.

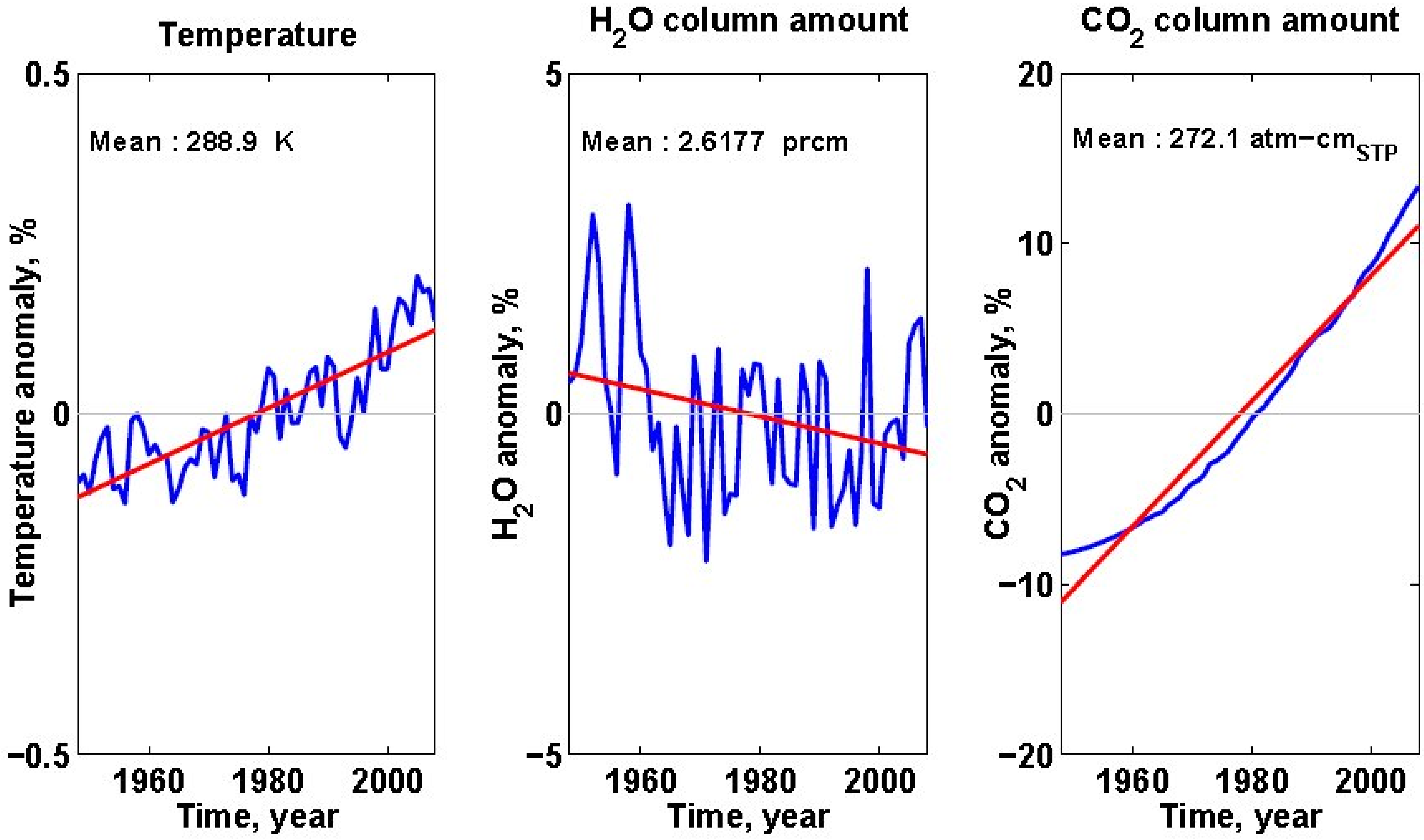


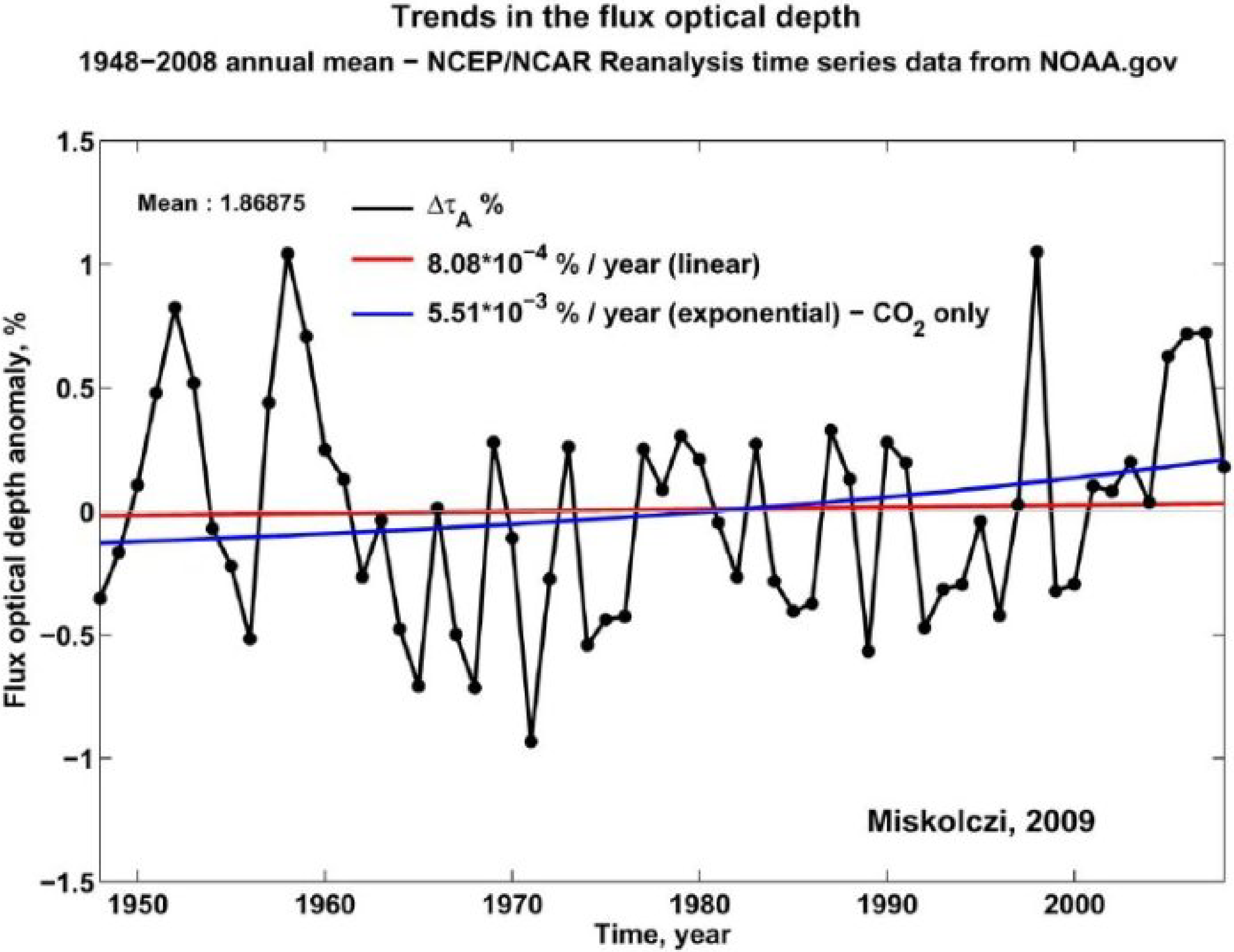
Interpretation (cont.):

- **If the system energetically could increase its surface temperature, it need not wait for our anthropogenic CO_2 emissions, since another GHG, water vapor, is available in a practically infinite reservoir, in the surface of the oceans.**
- **Energetic constraints can compensate the increasing CO_2-amount in the air for example by removing water vapor, rearranging its spatial distribution, or by modifying the amount (~62%) and/or the average height (~2 km) of the partial cloud cover.**

Linear trends in the NOAA 61 year time series (1948−2008)

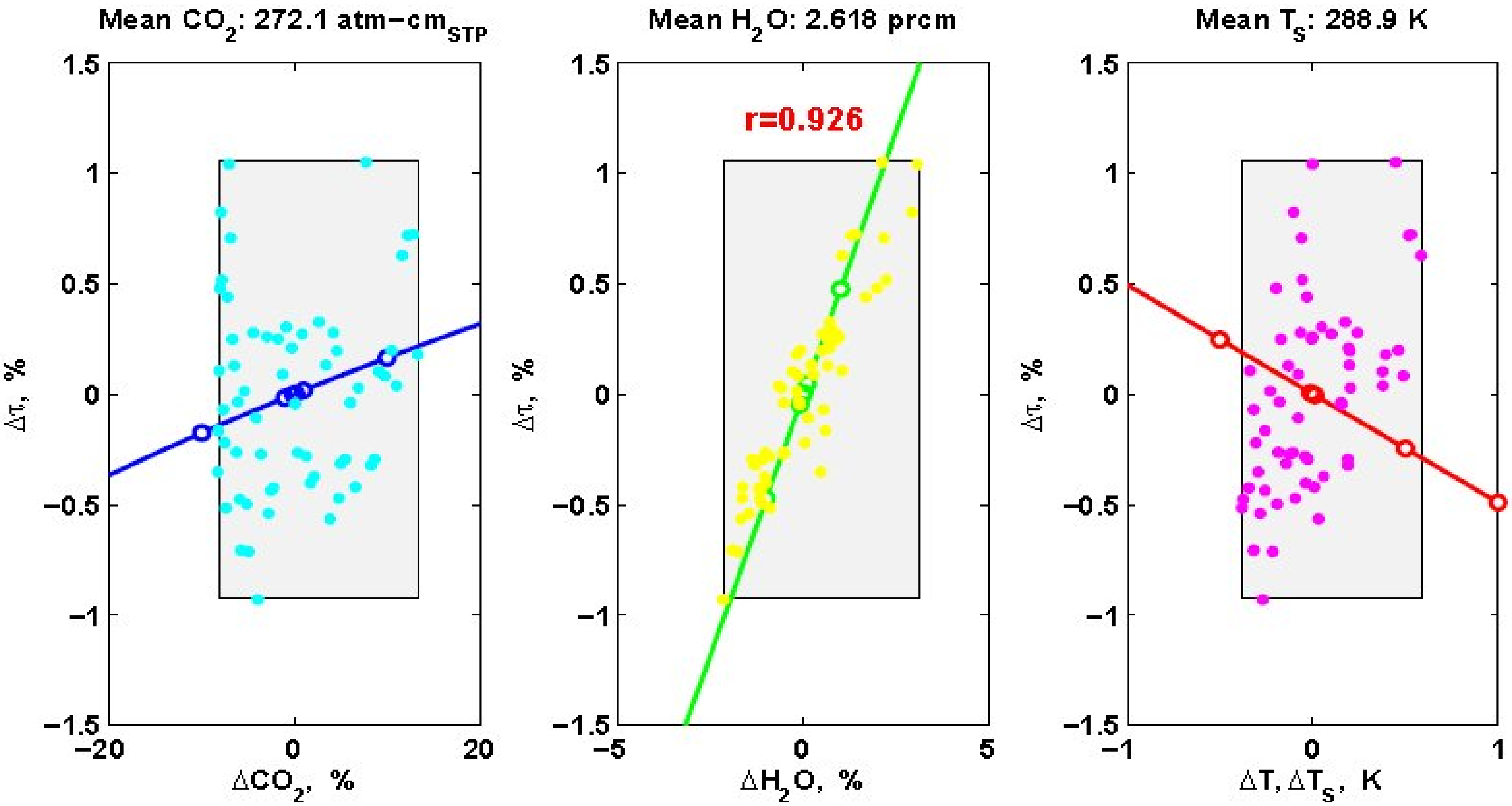
http://www.cdc.noaa.gov/cgi-bin/Timeseries




Trends in the flux optical depth
1948−2008 annual mean − NCEP/NCAR Reanalysis time series data from NOAA.gov
Mean : 1.86875
Δτ_A %
8.08*10^−4 % / year (linear)
5.51*10^−3 % / year (exponential) − CO_2 only
Flux optical depth anomaly, %
1.5
1
0.5
0
−0.5
−1
−1.5
1950
1960
1970
1980
1990
2000
Time, year
Miskolczi, 2009

HARTCODE input layer perturbation summary

Perturbations applied to each layer of the NOAA 61 year annual mean profile, mean τ : 1.86877



The dots in the gray areas indicate the true optical depth anomaly in each of the 61 years

Only the atmospheric H_2O can be related to flux optical depth anomalies

The previous graphs show

- a certain decrease in the integrated water vapor column amount in the ~1950-1970 time period,
- almost stable conditions during 1970-1990
- and, in good agreement with the literature beyond IPCC AR4 Chapter3, a slight increase in the past two decades (Trenberth, Santer, Willett, Wentz etc.)
- The whole process indicates the work of a com-pensation effect on a multidecadal timescale.

How much water vapor is needed to compensate the CO_2 effect?

The derivative of the

$$g(\tau) = G/S_U = (S_U - OLR)/S_U = 1 - f =$$
$$= (\tau - 1 + \exp(-\tau)) / (\tau + 1 + \exp(-\tau))$$

normalized greenhouse function gives the sensitivity of $g(\tau)$ to the optical thickness:

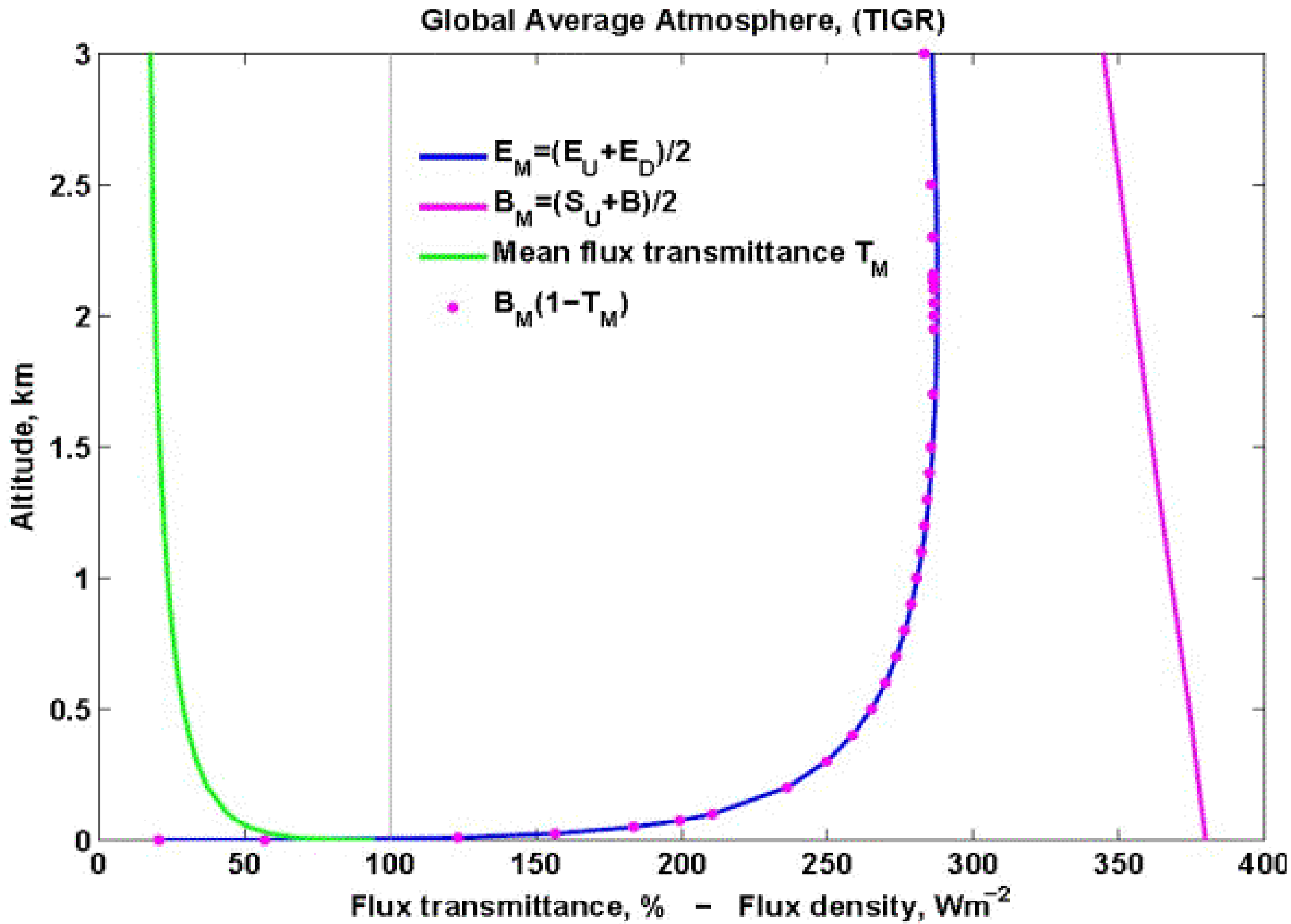
$$dg(\tau)/d\tau = 2(1 - \exp(-\tau)) / (1 + \tau + \exp(-\tau))^2 .$$

According to this, removing all CO_2 from the air would have an initial effect to diminish the surface temperature by 2.5K.

To compensate this, an increase of 0.08 prcm (3%) in the global average water vapor would needed. Or, a decrease of the same amount would completely hide the greenhouse effect of the CO_2 doubling. This analytical result compares very well with the detailed LBL calculations.

Some comments on the new equations

Eq. (4) $A_A = E_D$ works also in cloudy conditions:



Below the cloud layer there is radiative equilibrium, over the cloud there is clear-sky greenhouse effect.

Eq. (5) $E_U = F + K + P$ shows that

- the source of the upward atmospheric radiation is not related to LW absorption process;
- its zonal averages are almost independent of the water vapor column amount;
- have no meridional variation;
- is constantly the half of the surface upward flux density.

Eq. (8) $(S_U - OLR) + (E_D - E_U) = F^0 = OLR$

as noted, assumes the optimal (that is, maximal) conversion of SW into LW, that is, F^0 into OLR.

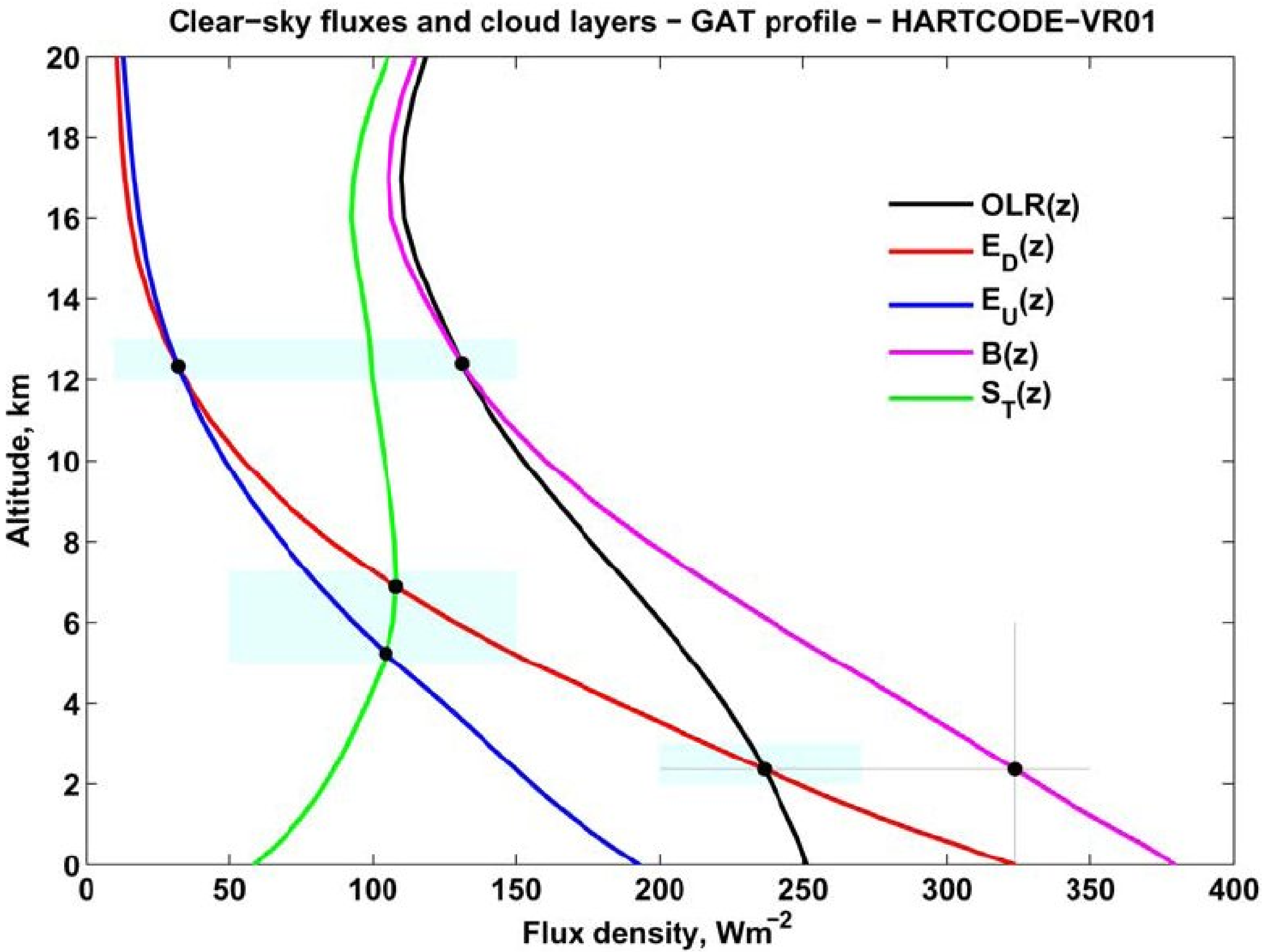
Its general form is

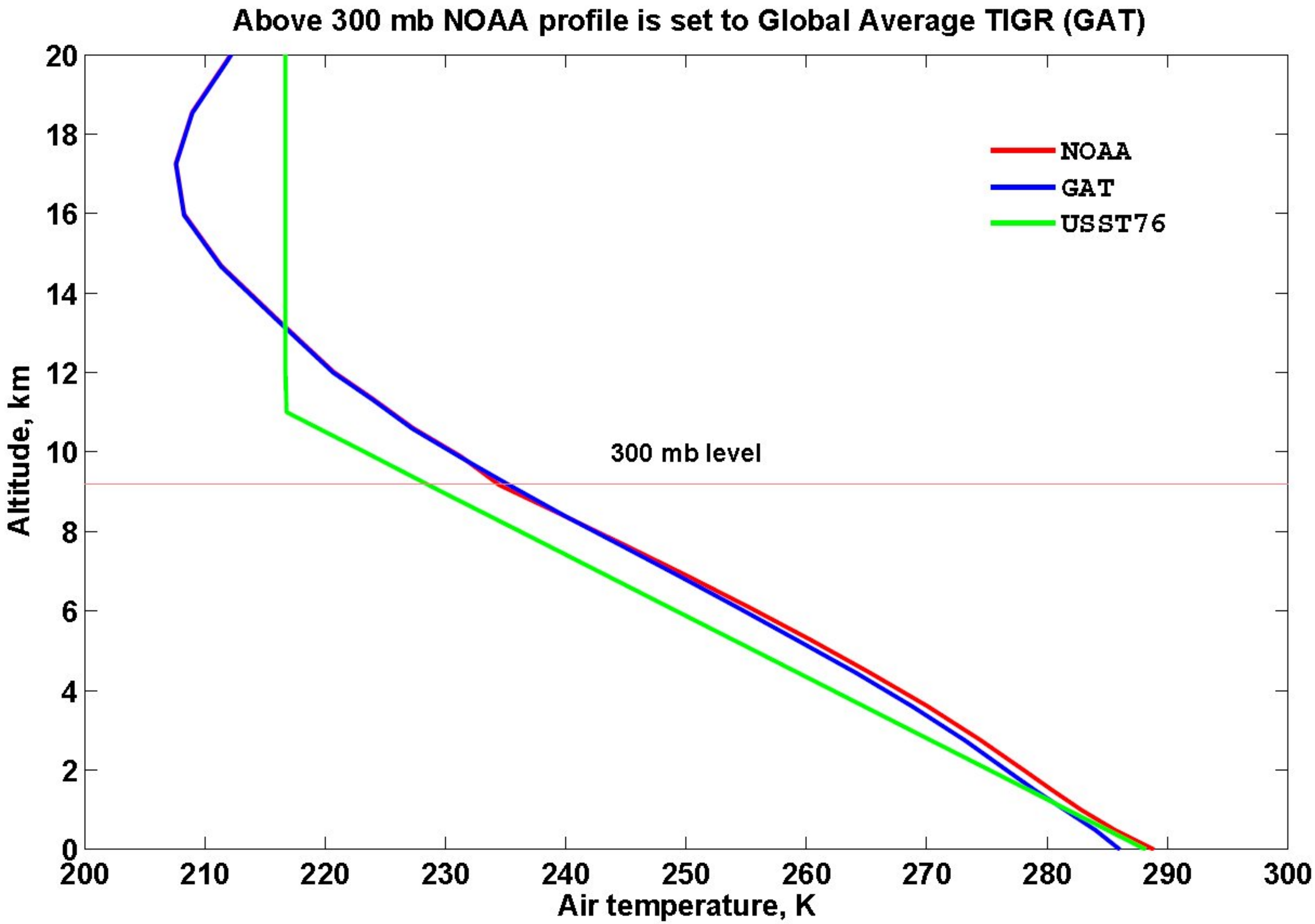
$$(S_U - OLR) + (E_D - E_U) = F^0 - S_T = OLR - S_T$$

* Moon: $S_T = S_U = OLR$, $E_D = E_U = 0$,

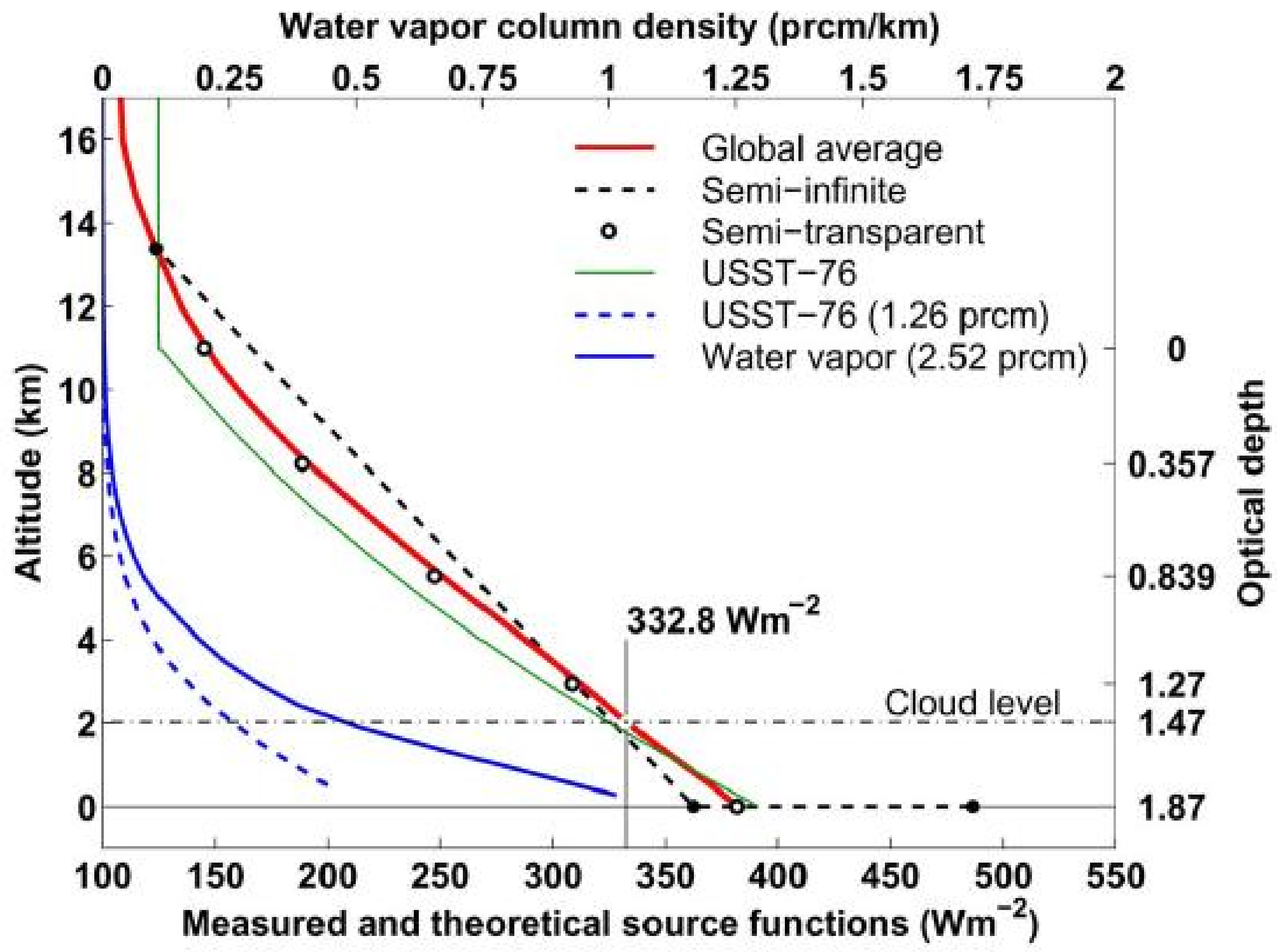
* Mars: $S_T/S_U=0.84$ (thin CO_2 atmosphere, no clouds)

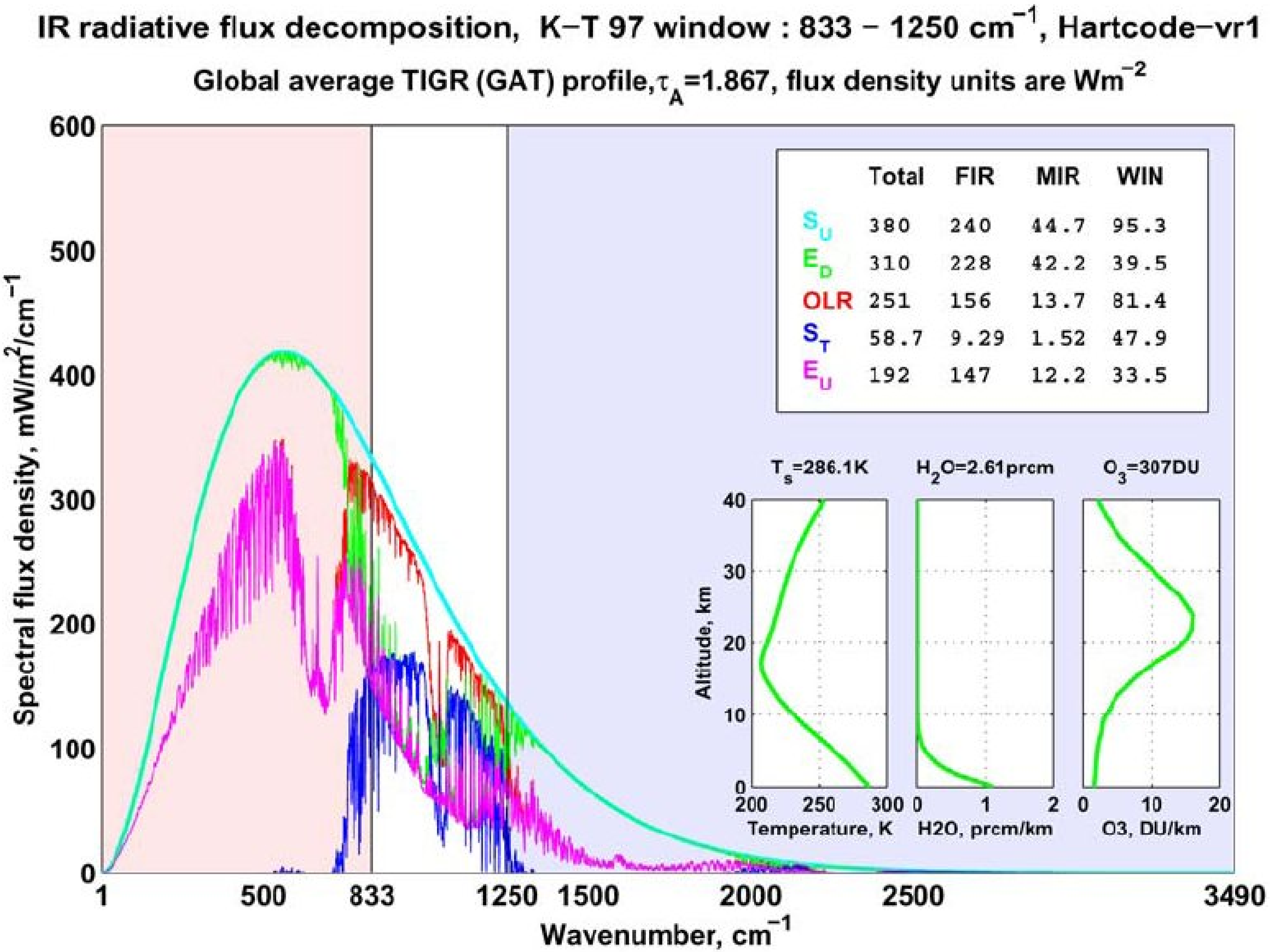
* Earth: the molecular structure of the GHG's does not allow S_T to be zero (there is "window" radiation). But the Earth's atmosphere has an extra degree of freedom: clouds. A 62% partial cloud cover at ~2 km has an LW effect about to "close" the atmospheric IR window.


Clear−sky fluxes and cloud layers − GAT profile − HARTCODE−VR01
Altitude, km
Flux density, Wm^{-2}
0
2
4
6
8
10
12
14
16
18
20
0
50
100
150
200
250
300
350
400
OLR(z)
E_D(z)
E_U(z)
B(z)
S_T(z)

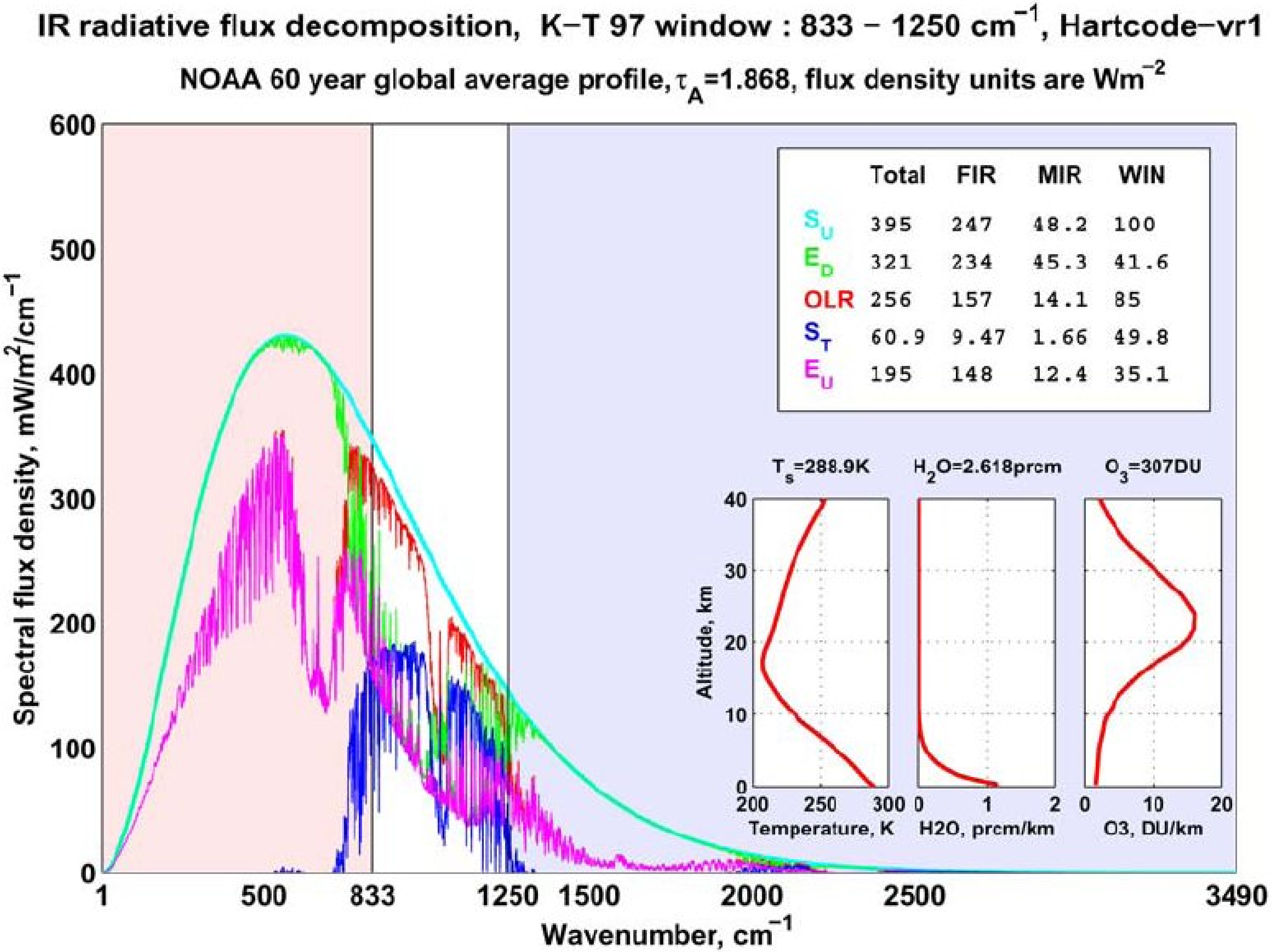

Above 300 mb NOAA profile is set to Global Average TIGR (GAT)
NOAA
GAT
USST76
300 mb level
Altitude, km
Air temperature, K
0
2
4
6
8
10
12
14
16
18
20
200
210
220
230
240
250
260
270
280
290
300

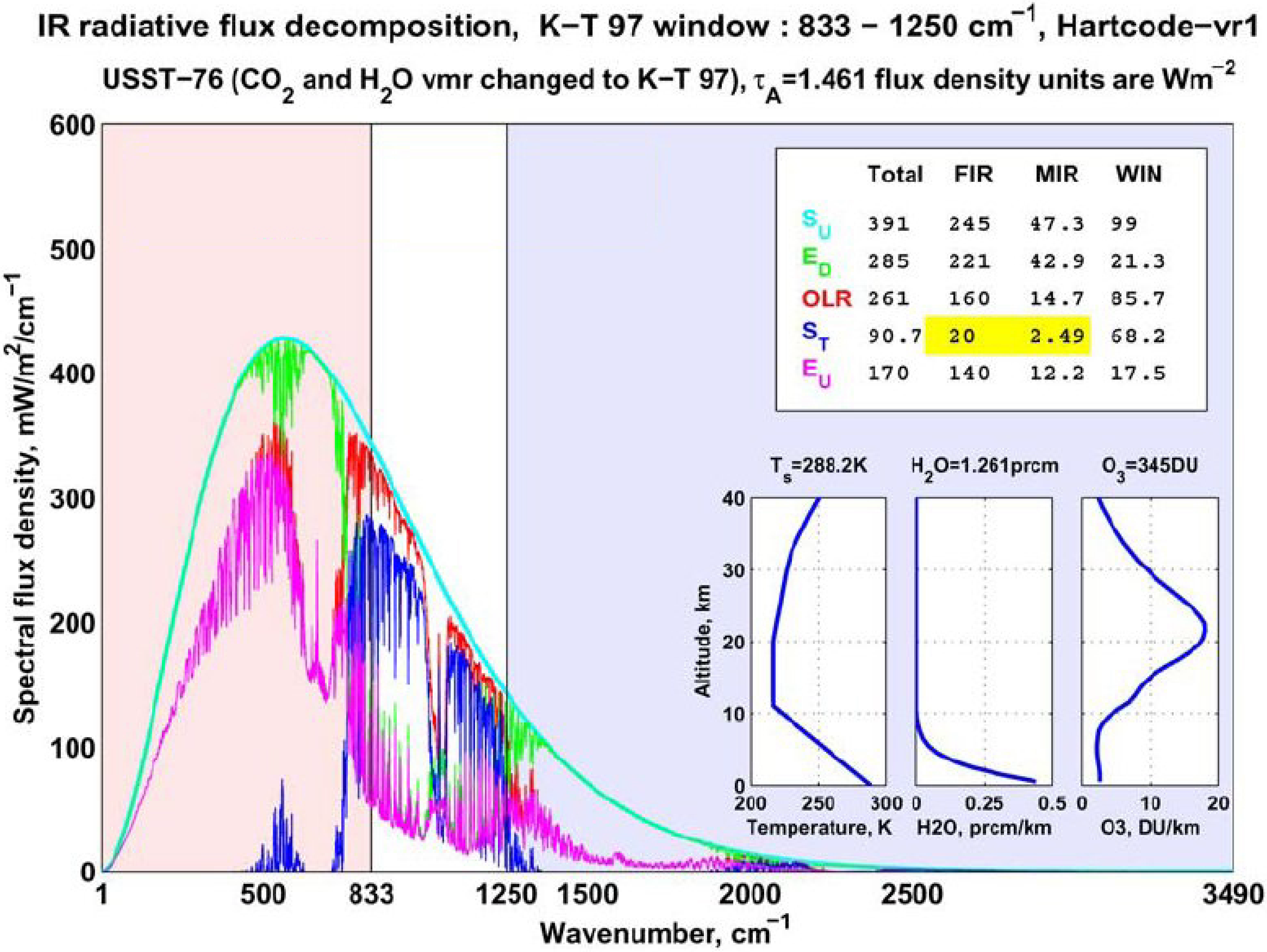
GLOBAL AVERAGE ATMOSPHERES





IR radiative flux decomposition, K−T 97 window : 833 − 1250 cm^{-1}, Hartcode−vr1
Global average TIGR (GAT) profile, τ_A=1.867, flux density units are Wm^{-2}
Spectral flux density, mW/m^2/cm^{-1}
Wavenumber, cm^{-1}
600
500
400
300
200
100
0
1
500
833
1250
1500
2000
2500
3490
Total FIR MIR WIN
S_U 380 240 44.7 95.3
E_D 310 228 42.2 39.5
OLR 251 156 13.7 81.4
S_T 58.7 9.29 1.52 47.9
E_U 192 147 12.2 33.5
T_s=286.1K
H_2O=2.61prcm
O_3=307DU
Altitude, km
Temperature, K
H2O, prcm/km
O3, DU/km


IR radiative flux decomposition, K−T 97 window : 833 − 1250 cm−1, Hartcode−vr1
NOAA 60 year global average profile, τA=1.868, flux density units are Wm−2
Spectral flux density, mW/m2/cm−1
Wavenumber, cm−1
600
500
400
300
200
100
0
1
500
833
1250
1500
2000
2500
3490
Total FIR MIR WIN
SU 395 247 48.2 100
ED 321 234 45.3 41.6
OLR 256 157 14.1 85
ST 60.9 9.47 1.66 49.8
EU 195 148 12.4 35.1
Ts=288.9K
H2O=2.618prcm
O3=307DU
Altitude, km
Temperature, K
H2O, prcm/km
O3, DU/km



IR radiative flux decomposition, K−T 97 window : 833 − 1250 cm−1, Hartcode−vr1
USST−76 (CO2 and H2O vmr changed to K−T 97), τA=1.461 flux density units are Wm−2
Total FIR MIR WIN
SU 391 245 47.3 99
ED 285 221 42.9 21.3
OLR 261 160 14.7 85.7
ST 90.7 20 2.49 68.2
EU 170 140 12.2 17.5
Ts=288.2K
H2O=1.261prcm
O3=345DU
Altitude, km
Temperature, K
H2O, prcm/km
O3, DU/km
Spectral flux density, mW/m2/cm−1
Wavenumber, cm−1
1
500
833
1250
1500
2000
2500
3490

Summing up:

in the Earth's atmosphere, equations

- $S_U = E_D/A \quad (A_A = E_D)$
- $S_U = 2E_U$;
- $S_U = (3/2)OLR$

describe *facts*,

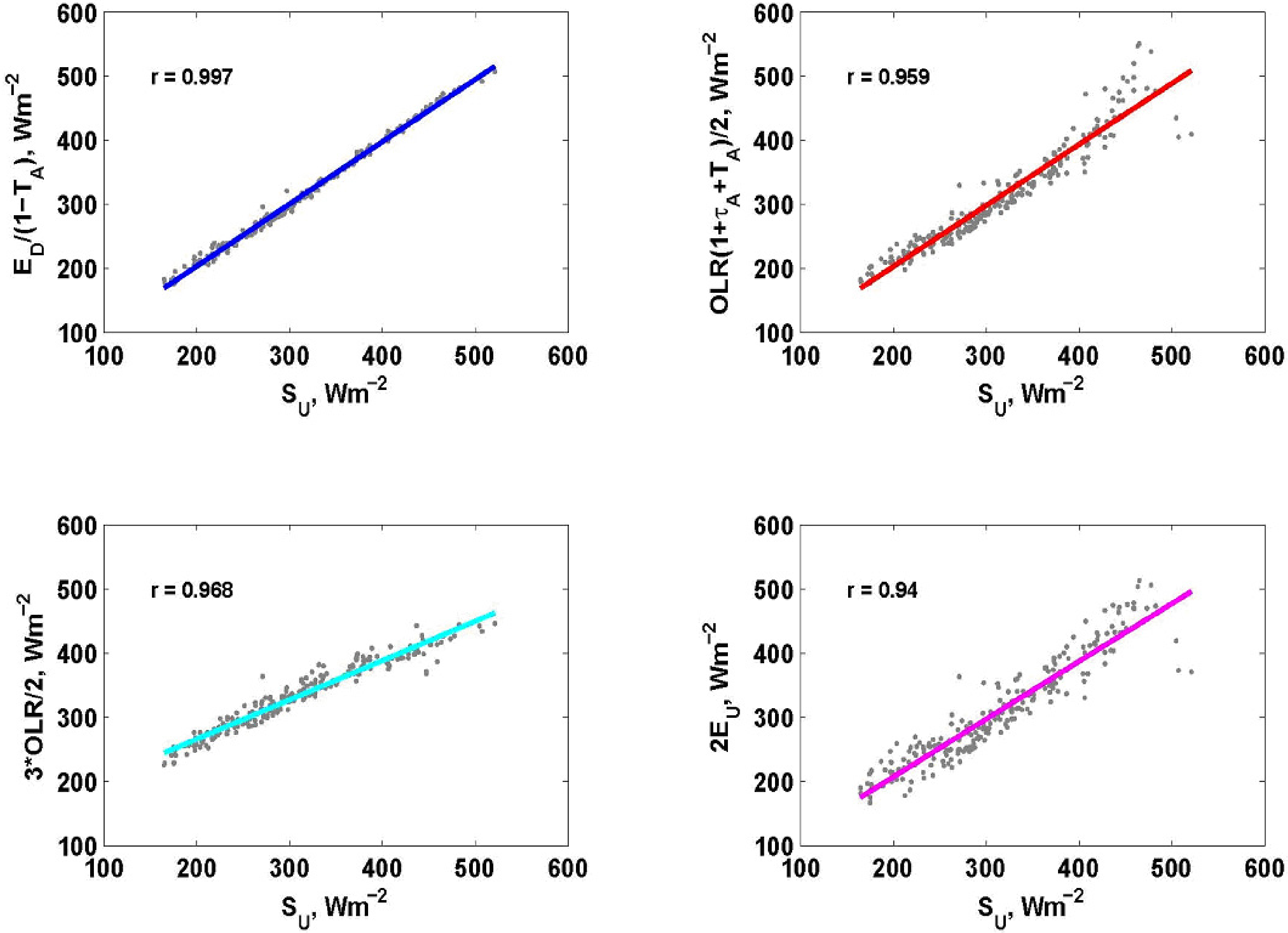
- $S_U = OLR/f \quad (f = 2/(1+\tau_A+\exp(-\tau_A))$

is a new proved *theoretical relationship*.

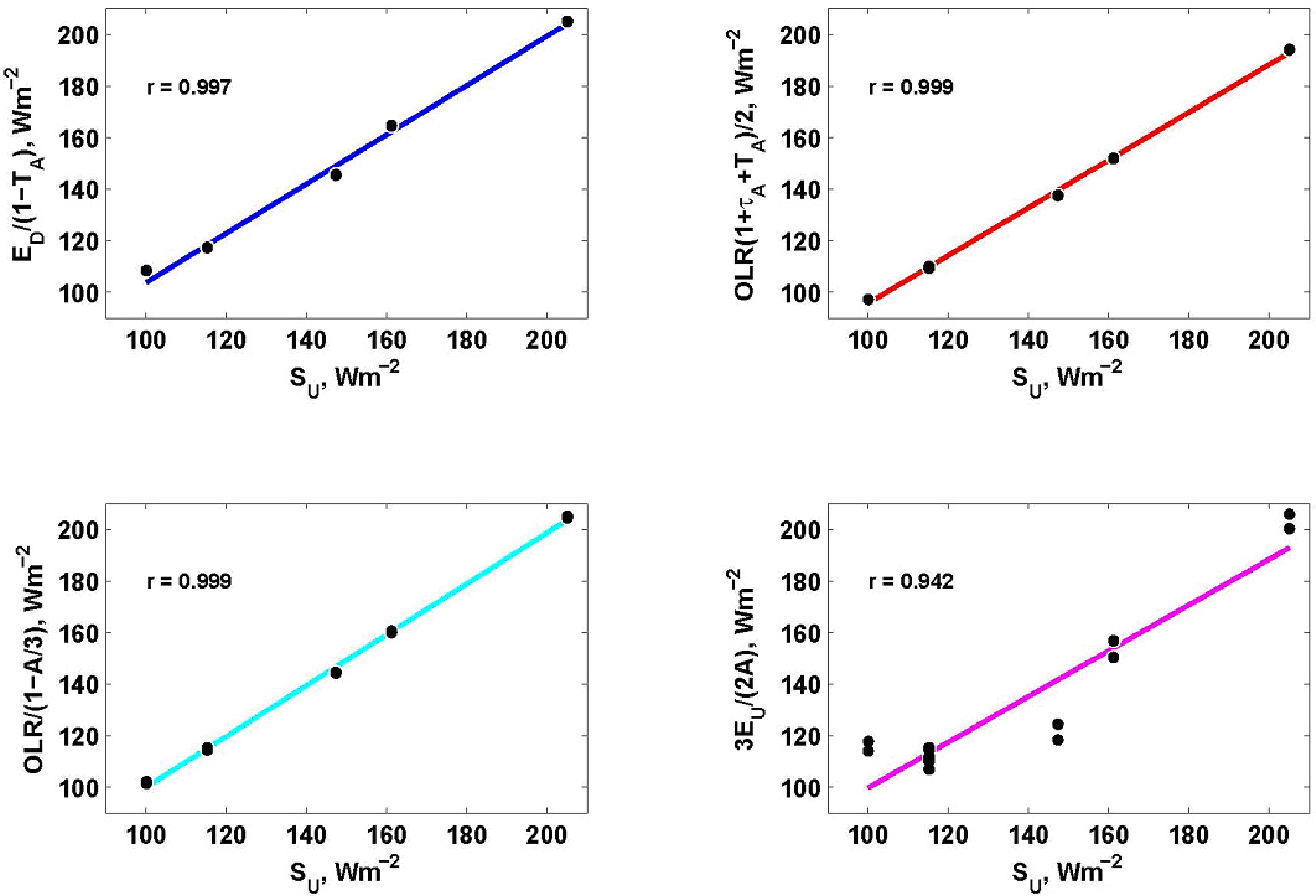
They all have to be taken into account in every valid annual global mean energy budget.

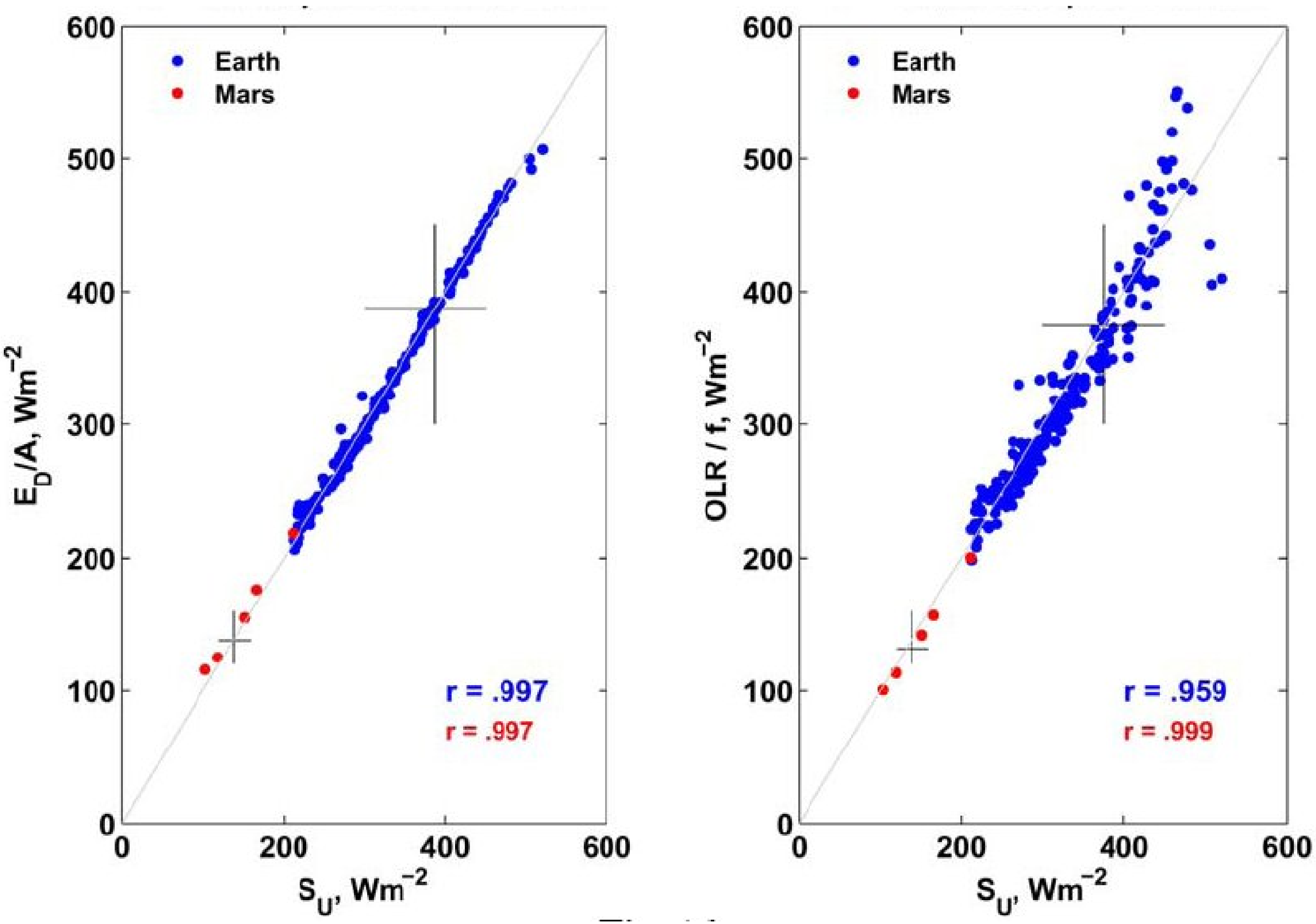
Their theoretical explanation (thermal equilibrium, radiative equilibrium, local thermodynamic equilibrium, Kirchhoff law, virial theorem, radiation pressure, entropy maximum, energy minimum, most effective cooling, the role of clouds, compensation mechanisms, equilibrium time scales, further planetary applications etc.) can be debated.

Earth Atmosphere

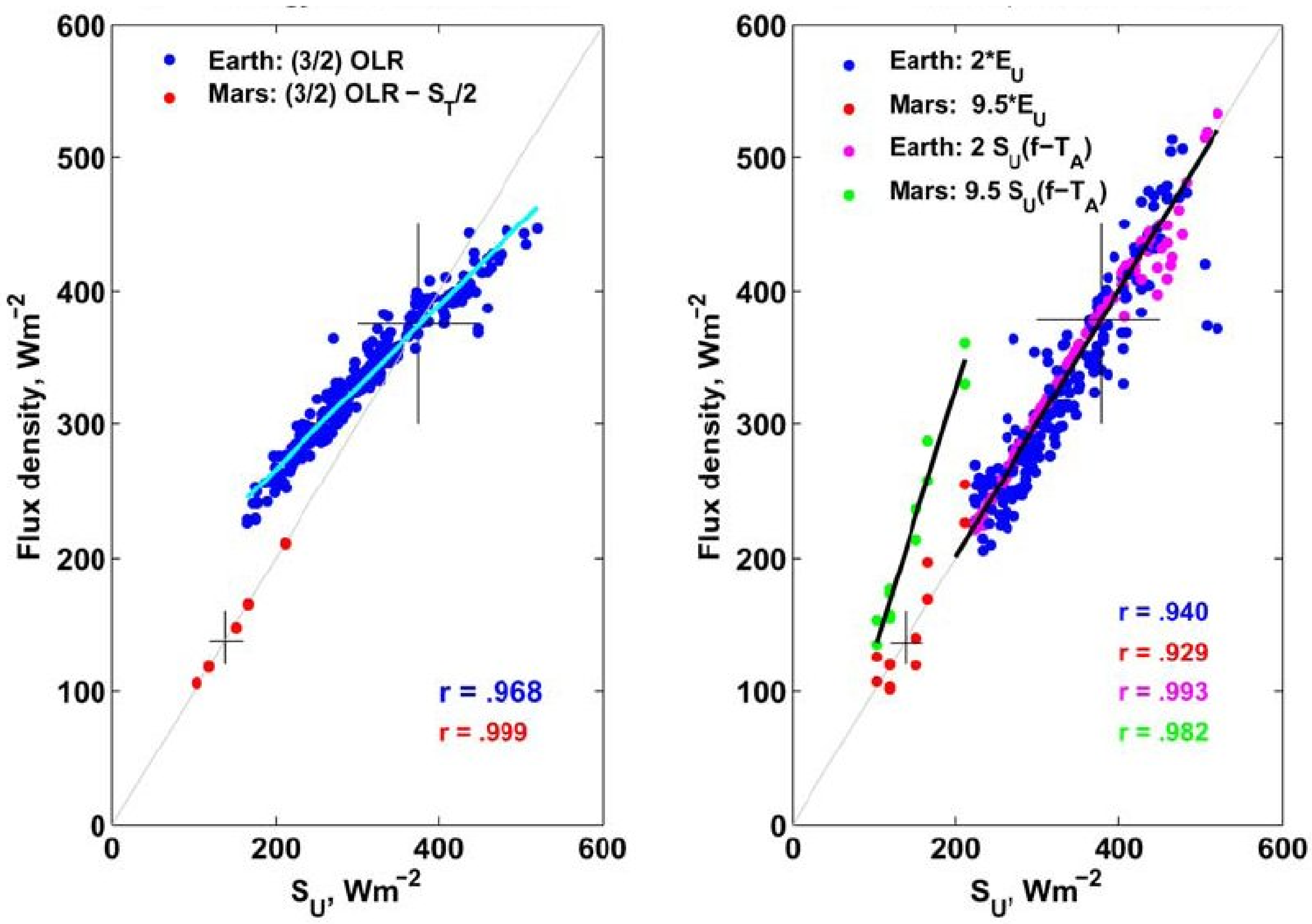


Martian Atmosphere



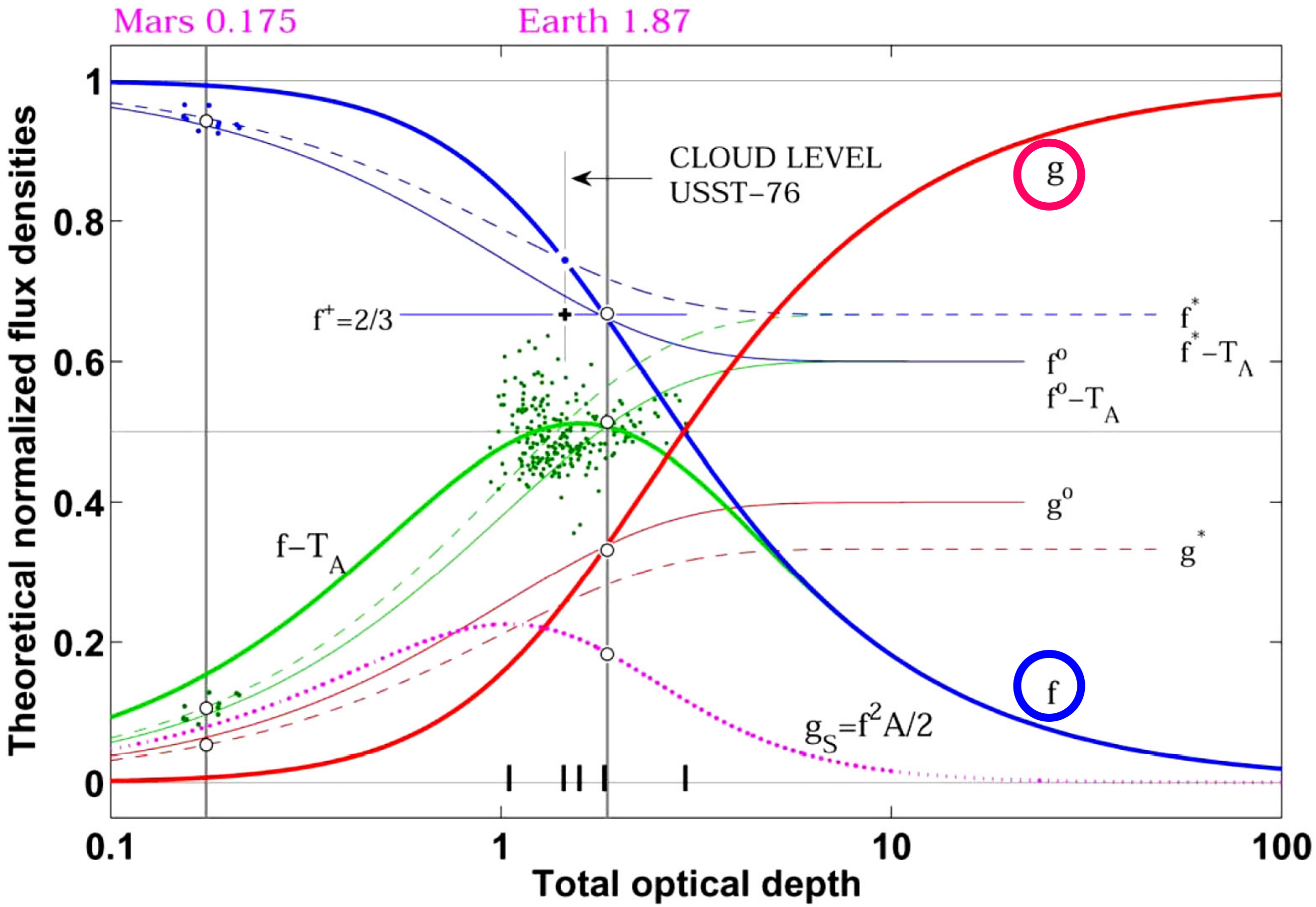


Earth
Mars
r = .997
r = .997
E_D/A, Wm^-2
S_U, Wm^-2
Earth
Mars
r = .959
r = .999
OLR / f, Wm^-2
S_U, Wm^-2



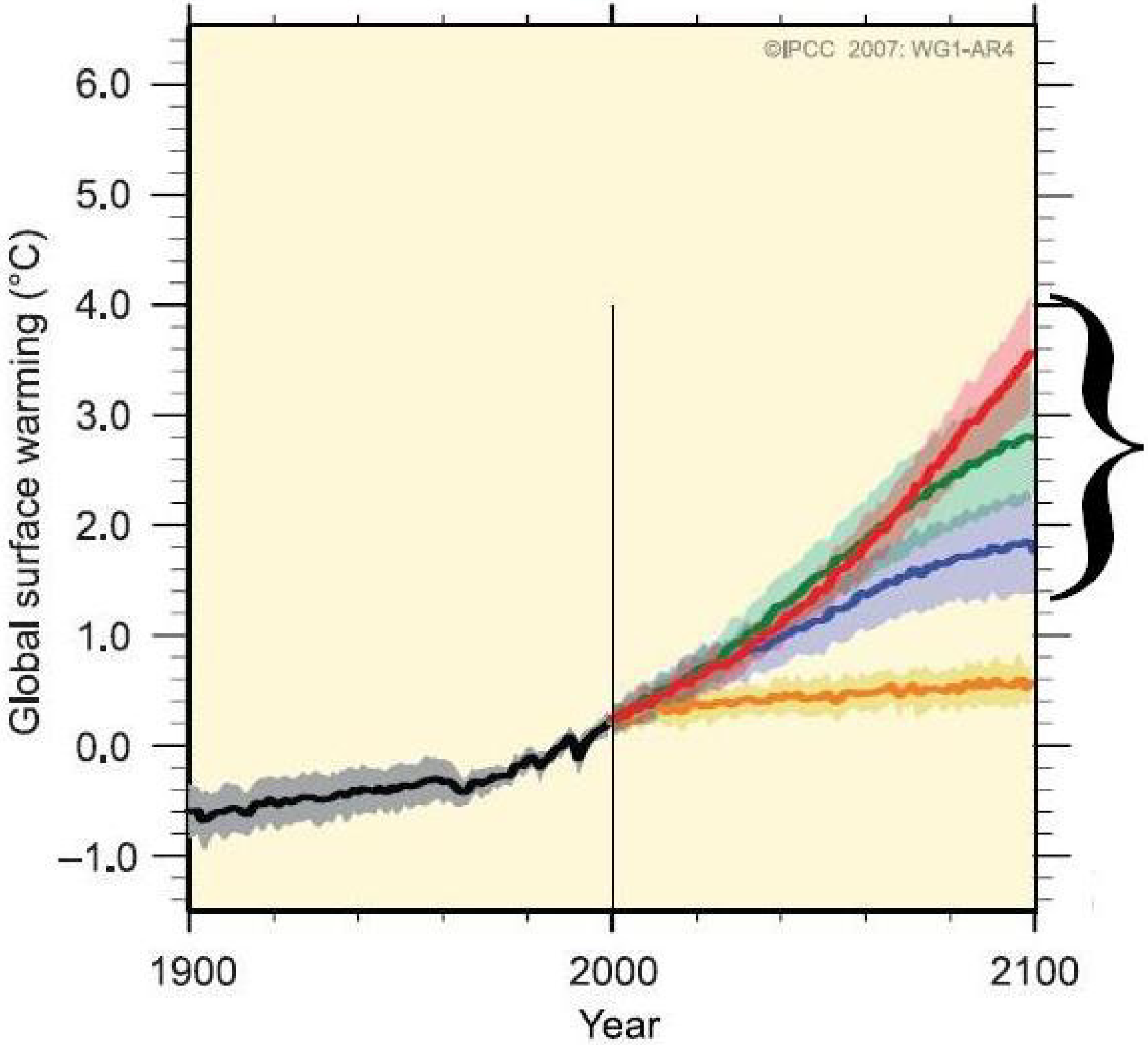
Earth: (3/2) OLR
Mars: (3/2) OLR − S_T/2
r = .968
r = .999
Flux density, Wm^-2
S_U, Wm^-2
Earth: 2*E_U
Mars: 9.5*E_U
Earth: 2 S_U(f−T_A)
Mars: 9.5 S_U(f−T_A)
r = .940
r = .929
r = .993
r = .982

GREENHOUSE EFFECT IN PLANETARY ATMOSPHERES



For the Earth and Mars the new theory is perfectly supported by simulation results and observations. The greenhouse effect on the Earth is locked to the critical optical depth of ~1.87 which is maintained by the atmospheric H_2O amount.

Temperature depends on
$F^0 + P^0$



Simply by emissions impossible.

Literature:

F. Miskolczi, *Időjárás, 2007, 111, No. 1* :
Greenhouse effect in semi-transparent planetary tmospheres.

F. Miskolczi and M. Mlynczak, *Időjárás, 2004, 108, No. 4* :
The greenhouse effect and the spectral decomposition of the clear-sky terrestrial radiation.

F. Miskolczi, *Időjárás, 2001, 105, No. 4* :
High accuracy skin temperature retrieval from spectral data of multi-channel IR imagers.

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miklos.zagoni@gmail.com

Part VII. Further considerations on the shortwave energy balance conditions

- In the longwave, we have seen that the balance of the two opposite regulatory powers led to a dynamic equilibrium value of the infrared greenhouse factor

$$G = S_U - OLR = F^0/2 \; (= OLR/2 = S_U/3) \; .$$

We suspect that the same must work also in the shortwave.

- The maximum entropy production principle requires the highest possible heating of the surface. For that reason, all the incoming SW radiation should reach the surface:

$$F^0 - F = S^0,$$

where $S^0 = S^C/4 = 342 Wm^{-2}$, with $S^C = 1368 Wm^{-2}$ solar constant, $S^0 = F^0 / (1 - \alpha)$ by definition, α is the planetary albedo, and F is the SW radiation absorbed in the atmosphere, $F \approx 67\ Wm^{-2}$.

- On the other side, the energy minimum principle prescribes that the downward SW radiation at the surface should be zero:

$$F^0 - F = 0$$

- The equilibrium value between the two is

$$F^0 - F = F^0/2(1 - \alpha) = S^0/2 = \text{const.} \qquad \text{(Z1)}$$

- F is determined overwhelmingly by the β global average cloud coverage ratio, the cloud albedo and the cloud SW absorptivity, but depends also on the SW absorption properties of the cloudless atmosphere (ozone, aerosols, diffusion, scattering etc.).
- From the approximate first order cross-dependences of F^0, F, α and β in Eq.(Z1), neglecting the non-cloudy components in F, we can derive a crude estimate for the equilibrium values, according the two opposite powers in the following form:

$$\alpha = \beta / 2 \qquad \text{(Z2)}$$

and

$$\alpha = (1 - \beta) / (2 - \beta) \qquad \text{(Z3)}$$

Eq.(Z2) describes the control from β=0 (α=0) to β=1 (α=1/2), while Eq.(Z3) from β=0 (α=1/2) to β=1 (α=0).

The solution of Eqs.(Z2-Z3) is
$\alpha = (2-\sqrt{2})/2$ and $\beta = (2-\sqrt{2})$ $(\alpha \approx 0.293 , \beta \approx 0.586)$,
while their observed global average values are $\alpha \approx 0.3$ and $\beta \approx 0.6$.
A better fit from this approximate description cannot be awaited.

FINAL SUMMARY

- The laws of physics relate unequivocally the longwave greenhouse factor, G, to the incoming available shortwave energy, F^0.
- The same principles determine the equilibrium value of the available surface shortwave energy, $F^0 - F$, as a function of the solar constant S^C or S^0, through controlling the planetary albedo α and the fractional cloud cover β.


Thank you.