Tananyag fejlesztés idegen nyelven

Prevention of the atmosphere

KÖRNYEZETGAZDÁLKODÁSI AGRÁRMÉRNÖKI MSC

(MSc IN AGRO-ENVIRONMENTAL STUDIES)

Új Magyarország Fejlesztési Terv

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Fundamentals to global warming (radiative forcing, feedback factor, GWP, LOSU)

Lecture 9

Lessons 25-27

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Lesson 25

Solar radiation types. Radiation absorption in the air – consequences. Passive remote sensing. Radiation balance (short wave)

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- Atmospheric scientist distinguish two parts of the solar radiation:

  - The short wave or solar radiation is in the range of 0.15 – 3.0 μm (energy emission of the Sun) – Fig. 66.

  - the long wave radiation is in the range of 3.0 – 15 μm (terrestrial radiation)

The energy conservation states the way of incident radiation in a medium (as in the air also); it may

- absorb
- transmit or
- reflected from the given surface.

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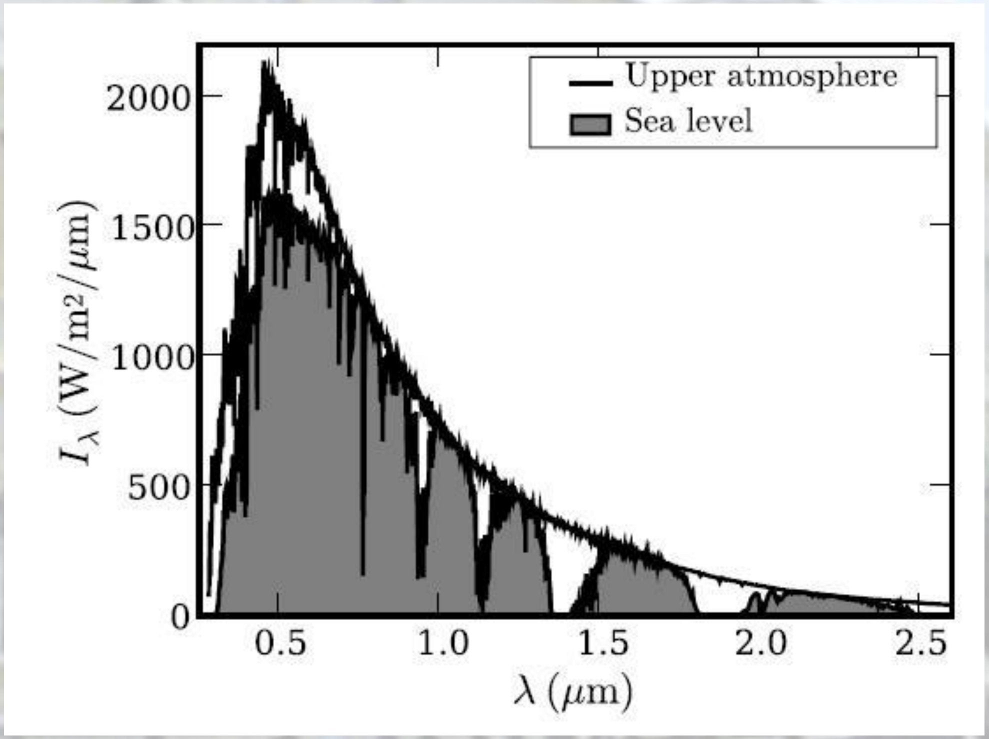
Fig. 73 Short wave radiation at two levels (after Sportisse)



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Radiation absorption

The absorption for UV radiation is due to the presence of molecular oxygen, ozone, water vapor and carbon dioxide.

- The short wave radiation is absorbed in the ionosphere, mesosphere and stratosphere.

The atmopsheric window is the place where the atmosphere is transparent to light. The consequence is the heating of the surface of the Earth.

The longwave radiation (resulted from emissions) is absorbed by water vapor, methane, CO_2 and ozone.

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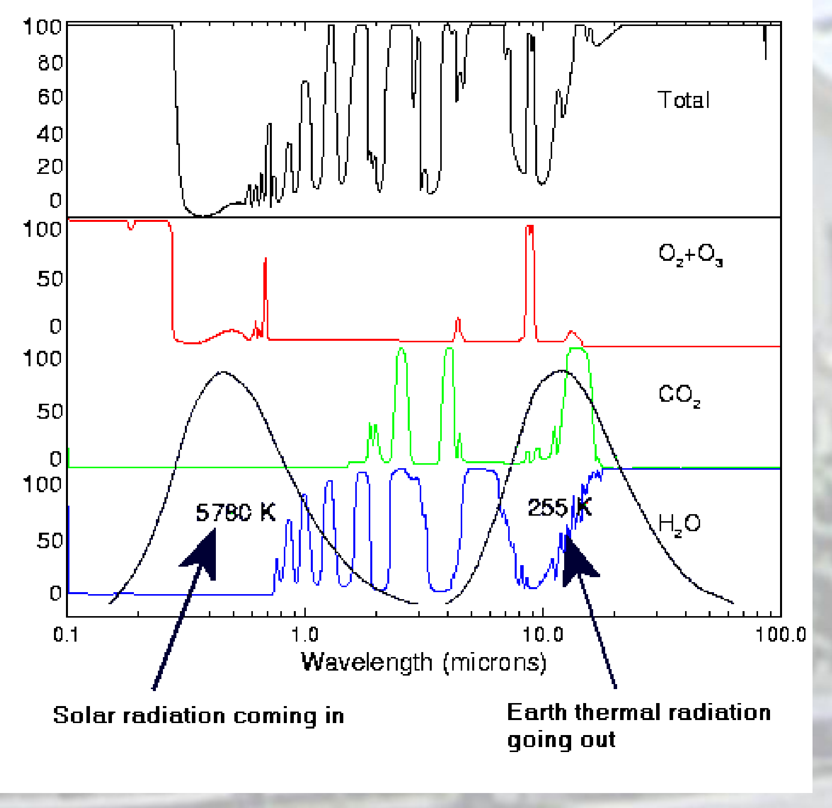
Fig. 74 Infrared Absorption Spectrum for CO_2 and some other greenhouse gases



www.te-software.co.nz/blog/augie_auer.htm

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These gases absorb in the infrared region of radiation. They are responsible for greenhouse effect.

Passive remote sensing

This phenomenon is applied by the satellite platforms.

The basis is the absorption of the atmosphere in the infrared radiation region.

Measured spectrum by the satellites indicate the vertical distribution of atmospheric element, mainly the water vapor and the temperature.

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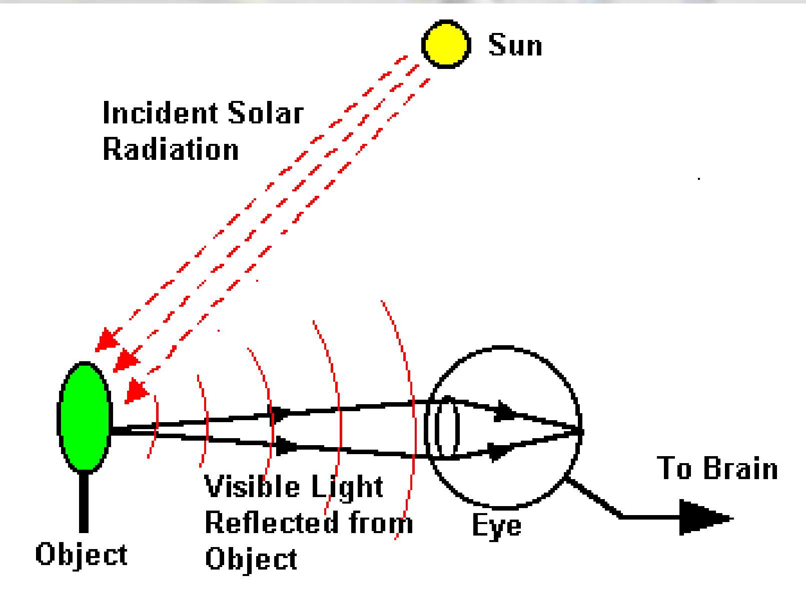
Fig. 75 Passive Remote Sensing – for eyes



The eyes passively sense the radiation reflected or emitted from the object. The sensing system depends on an external source of illumination.

www.physics.nus.edu.sg/.../tutorial/eye.htm

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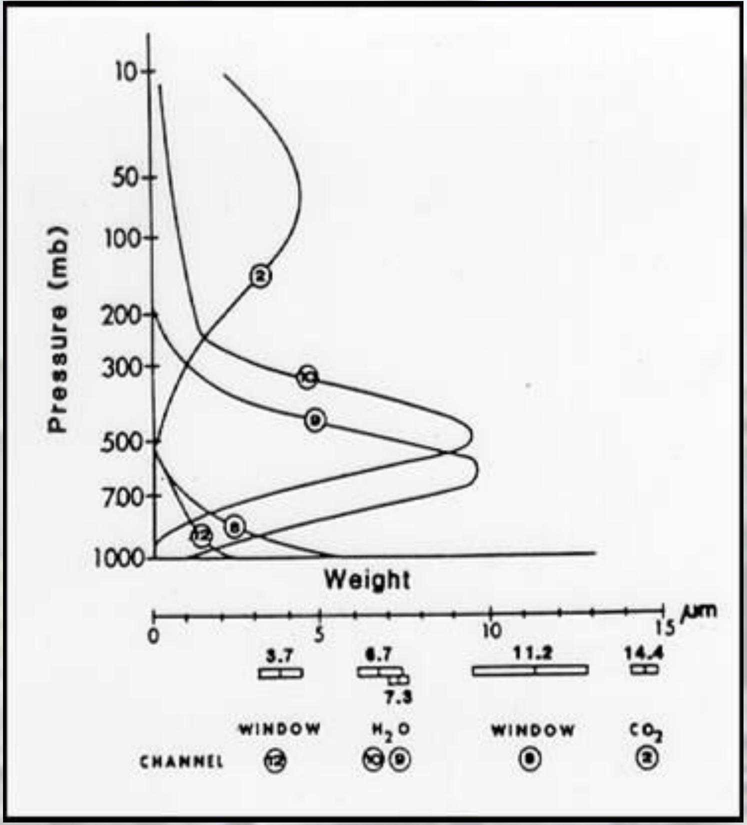
Fig. 76 Passive Remote Sensing – for satellites



www.saao.ac.za/.../ESOph5/UseSat_InvPap_Mar.html

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Fig. 76 – discussion 1

Measurements of water vapor and cloud are carried out by meteorological satellites by passive remote sensing at different wavelengths. Depending on the wavelength of the emissions being measured by the satellite, different quantities can be derived. Figure shows the weighting functions for different infrared channels. In the IR window channel (10.7μm), for example, emissions reach the satellite essentially unattenuated by the atmosphere so that radiance values measured are due to emission from the surface.

Figure 76 - discussion 2

It indicates, that observations at 6.7μm are sensitive to emission by water vapor in the layer between 600mb and 300mb. (There are only small amounts of water vapor above 300mb.) Emission from this layer depends on the amount of water vapor in the layer and temperature. Temperature can be accounted for by using an observed or representative temperature-height sounding so that emission is then only a function of the amount of the emitting gas, in this case water vapor, in the layer. So the water vapor can be observed properly.

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- The sum of the three components as expressed in forms of absorptivity (abs), transmittance (τ) and reflectivity (α) is always equal to one in a given moment:

$$abs + \tau + \alpha = 1$$

Radiation balances are exist in the Earth-atmosphere system.

Part of the short wave radiation (G) reflects into the space determined by the surface albedo (a). The net game (I) will be:

$$I_{inc} = G - aG = G\,(1-a)$$

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Lesson 26

Radiation balance (long wave). Net radiation. Energy balance of the Earth-Atmosphere system

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The outgoing radiation (I_{out}) of the Earth is governed by its surface temperature (T) as controlled by the Stefan-Boltzmann law:

$$I_{out} = \varepsilon \sigma T^4 + (1-\varepsilon) Le$$

where ε is the emissivity (less than unit)

Le is the amount of long wave radiation reflected

σ is the S-B constant

The net all wave radiation for the surface is the difference between the above two balances, the so called the net radiation (R_{net}):

$$R_{net} = I_{inc} - I_{out}$$

The net all wave radiation is the basis in energy exchange processes, because it determines the available energy for the consumers to energy balance:

$$R_{net} = Q_E + Q_H + Q_S + Q_P$$

Where Q_E stands for the latent heat (evaporation)

Q_H the sensible heat

Q_S the soil heat flux

Q_P energy bound in photosynthesis

The mean values (whole world) are about 70%, 15%, 15% and less than 1% for Q_E, Q_H , Q_S and Q_P, respectively.

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Energy balance of the Earth-Atmosphere system

The Earth-Atmosphere system consists of three sub-systems:

- the Earth
- the Atmosphere and
- the Space.

In annual basis the whole system is in equilibrium. It is a closed system for the mass transfer, but it allows an energy exchange to the space.

The input of the system is the solar radiation. The top of this energy is given by solar constant (*Io*).

The solar constant is the amount of radiation energy on unit surface placed normal to the actual solar beam on the top of the atmosphere. Its yearly mean is about 1370 W m^{-2}.

The exact solar constant is influenced by the actual position of the Sun and Earth. Its variability is not very high, it is close ± 3 to ± 4%. The annual mean value of radiation income is $I_o/4$ = 343 Wm^{-2} (29.7 MJ m^2 day^{-1}).

The solar constant is the starting point (upper limit), the 100% when discussing the energy balance of the whole Earth-Atmosphere system.

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Exercise

Control the annual variability of the solar constant, I_o!

The extreme distances (Earth-Sun) are

r’: 1.469 x 10^8 km

r”:1.520 x 10^8 km

Using the Stefan-Boltzmann law to Sun (its radius, R is 6.96 x 10^5 km; emission temperature, T of the Sun is 5783 K) the flux reads

$$I = \frac{R^2}{r^2} \sigma T_s^{\ 4}$$

The calculated fluxes will be

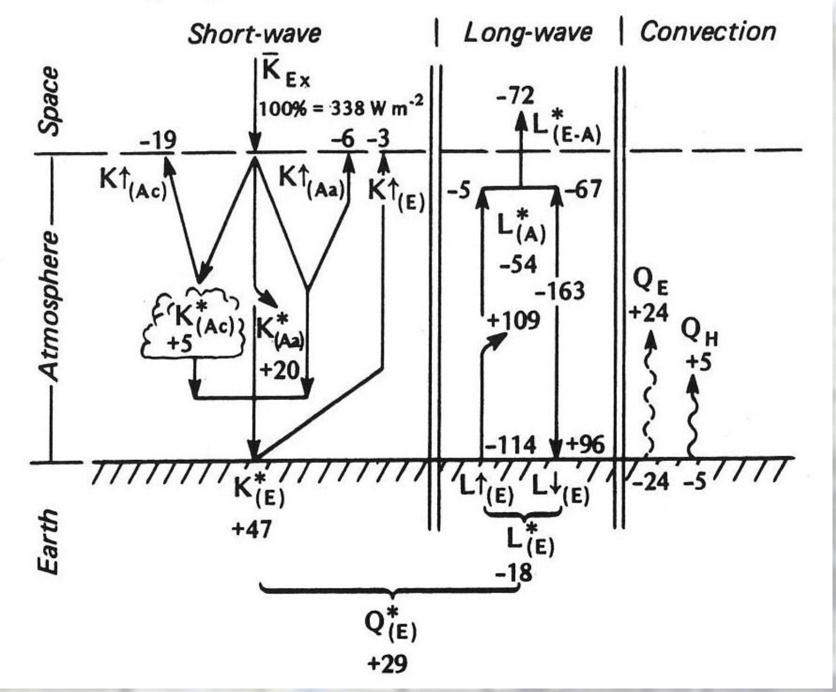
- The mean value: 1368 Wm^{-2}
- Closest Earth position to the Sun: 1410 Wm^{-2}
- Most far the Earth from the Sun: 1320 Wm^{-2}.

The difference between the two extreme positions is 90 Wm^{-2}.

Variability is only 5.7% comparing to the yearly mean of solar constant.

- As there is no shift in air temperature of the atmosphere, there is no gain or loss in energy on yearly basis. If this were not so, warming or cooling would be experienced in global air temperature of the Earth.
- The sub-systems of the Earth-Atmosphere are also in equilibrium. There is a special method to achieve this state that is working in both system and sub-system levels.
- The evidence for equilibrium is that the solar input (100%) comes from the sum of short wave scattering and reflection and long wave emission of the Earth and the atmosphere:

Fig. 77 Energy balance of the Earth-Atmosphere



- In numbers (see Fig. 77):

$$[6 + 19 + 3] + [67 + 5] = 100$$

The sub-systems are not in equilibrium. The Earth gets 47% radiation from short wavelength, but the net loss of long wave radiation is 18%.

The Earth emits 114% to the air, and gets back 96%.

- The net all wave radiation for the Earth sub-system is positive, it is equal to minus 29% (25-54)!
- Due to this number the Earth has a yearly radiation surplus energy of 29%; and the same appears as a minus in case of the atmosphere.

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Lesson 27

The components of the heat balance. The radiative forcing (values). Climate sensitivity parameter. GWP

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Earth energy budget

- This sub-part, the Earth energy balance is connected to
  - Radiative energy
  - Sensible heat
  - Latent heat
  - Soil heat conduction

Earth absobs 168 Wm^{-2} in shortwave and 324 Wm^{-2} in the longwave radiation. It emitts 390 Wm^{-2}. It yields gain (102 Wm^{-2}) of radiative energy of the Earh.

In the end this energy is applied for non-radiative processes; for water evaporation and vertical air motion.

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Energy budget for the atmosphere

- The atmospheric absorption in the short wavelength is 67 Wm^{-2}. The same for long wavelength is 350 Wm^{-2}. The terrestrial emission (loss) is 519 Wm^{-2}.
- The final radiation budget for the atmosphere is negative, as:

$$350 + 67 - 519 = -102\ Wm^{-2}.$$

As an equilibrium state is exits, this loss is compensated by the latent and sensible heat fluxes coming from the Earth surface.

The total radiation absorption of the both sub-systems is 235 Wm^{-2}.

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- In spite of the mentioned energy disparity, there is no warming or cooling in the substances, neither in the Earth or in the atmosphere.

Energy fluxes

- The extra energy is transported by convection offsetting the energy deficit. The energy amount of sensible heat is 5%; and of the latent heat is 24%.
- It can not be followed on our sample Fig., because over a yearly time period the net sub-surface energy storage is equal to zero. Due to this phenomena the Fig. 77 does not contain the convection.

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Basic definitions to global warming

Radiative forcing after IPCC Third Assessment Report:

„ The radiative forcing of the surface-troposphere system due to perturbation in or the introduction

of an agent (say, a change in greenhouse gas concentrations) is the change in net

(down minus up) irradiance (solar plus long-wave; in Wm^{-2}) at the tropopause AFTER allowing

the stratospheric temperatures to readjust to radiative equilibrium, but with surface

and tropospheric temperatures and state held fixed at the unperturbed values."

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Let’s change the albedo (*α*) value, and determine the direction of modification! The incident solar radiation is F_s, the radiative forcing is *F.*

- First step - assuming an equilibrium state in radiative forcing:

$$F = F_s (1- \alpha) - \sigma T^4 = 0$$

Second step modification of the albedo (*α’*) perturbs the radiation also:

$$\Delta F = - F_s \alpha'$$

If increasing the albedo value, negative feedback is waited in radiative forcing (cooling)*.*

- The impact of radiative forcing on the temperatures (surface temperature is included) may be estimated by adding the radiative forcing to the solar energy income.

The values of radiative forcing are given in Table 10.

The *climate sensitivity parameter* (λ) can be got by knowing the radiative forcing as follows:

$$\lambda_o = \frac{\Delta T}{\Delta F}$$

It's dimension is K $(Wm^{-2})^{-1}$

Table 10 Calculated radiative forcing

Species X_i	ΔF_{X_i} ($W m^{-2}$)	Uncertainties	LOSU
Greenhouse gases	2.43	10%	High
incl. CO_2	1.46	–	–
incl. CH_4	0.48	–	–
incl. N_2O	0.14	–	–
incl. halogens	0.34	–	–
Stratospheric O_3	−0.15	67%	Medium
Tropospheric O_3	0.35	43%	Medium
Sulfate aerosols (direct)	−0.4	[−0.8, −0.2]	Low
Biomass burning aerosols (direct)	−0.2	[−0.6, −0.07]	Very low
Soot (elemental carbon, direct)	0.1	[0.03, 0.3]	Very Low
Organic aerosols (direct)	−0.1	[−0.3, −0.03]	Very Low
Mineral aerosols	[−0.6, 0.4]	–	Very Low
Indirect effect of aerosols	[−2, 0]	–	Very Low
Condensation trails (aircrafts)	0.02	350%	Very Low
Cirrus formation (aircrafts)	[0, 0.04]	–	Very Low
Surface albedo (land use cover)	−0.2	100%	Very Low
Solar activity	0.3	67%	Very Low

IPCC Third Report (2001)

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In our table we consider the influence of increased gas or particle on the outgoing energy flux. The possible feedbacks were neglected (simplified). The sizes of radiative forcing are very uncertain. It can be seen in brackets. LOSU means the level of scientific understanding (knowledge).

From the industrial revolution until the turn of the century the radiative forcing was about two and a half Wm^{-2} resulting an increment of 0.6 K in mean air temperature.

The climate sensitivity parameter is 0.25.

Feedbacks in the forming of our climate (global warming)

The impact of the „other" components in warming process is due to radiative forcing and temperature change. We have to take into account them when analyzing the climate modifications.

- Water vapor – the warmer the air temperature, the higher the vapor holding capacity of the air, that favors the gas-phase water in the air.

The water vapor is an important gas increasing the greenhouse effect. The water vapor serves as a positive feedback.

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- Albedo – decline in ice cover decreases the albedo and the amount of received radiation is increasing. (The reason of accelerated ice melting is the warming itself.) Also a positive feedback.
- Clouds and aerosols

Not as simple as the previous factors were. The effect of aerosols is not completely known. Much uncertainities can be found regarding the aerosols.

The clouds have two faces:

- they absorb the infrared radiation, and at the same time they also reflect them.

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Global warming potential (GWP)

The global warming potential is in close connection with residence time of different materials (gases and aerosols).

For a given gas, the *global warming potential* is the result of comparison over a time period, when the radiative forcing coming from an extra 1 kg emission of this gas with that of producing from the same emission of a reference gas (CO_2).

In general this reference gas used to be the carbon dioxide.

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- The methane's GWP is 23 times higher than that of the reference CO_2 gas. The same for freons – depending on their species - may be more than 1000 times in comparison to CO_2!
- There are two types of global warming potentials:

  - Direct GWP – the above definition belongs to this category

  - Indirect GWP – physical, chemical and other influences are also taken into account.

This latter category is very complex, it is not easy to discuss.

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Thank you for attention!

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