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The Influence of Carbon Dioxide Pollutants on the value of Outgoing Longwave Radiation of Earth-Atmosphere System

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Abstract

The presence of pollutants in the atmosphere such as carbon dioxide leads to the variation of the albedo value of the unique Atmosphere - Earth system. the application of the thermodynamic calorimetric expression allows obtaining of the empirical expression containing such parameters as albedo, temperature and the specific thermal capacity of the components of this unique complex Earth-atmosphere system. it is possible from this dependence to found the emission thermal capacity of this unique system. A difference can be observed as large as possible between the equilibrium temperature and of the effective temperature for higher values of albedo. this difference in temperatures is inversely proportional to the values of the specific thermal capacity. Another important moment is that the increasing of the atmospheric temperature results in a decrease of the effective temperature. The values of the wavelengths of outgoing longwave radiation increase with the increasing of albedo values and is of the order of 10 microns for minimal value of albedo and of the order 14 microns for the maximal one. The cooling of Earth's upper atmosphere by the increasing of the pollutants gases is explained quantitatively by the suggested quantitative expression.

Keywords: *Earth's atmosphere, oxygen depletion, greenhouse gases, accumulation of carbon dioxide, effective temperature, specific heat capacity, albedo, specific Earth's heat content, outgoing longwave radiation.*

Introduction

The composition of Earth's ground together with its atmosphere in general is very complex by chemical structure point of view and by macro components of the Biosphere in general. [1]

The atmosphere and the solar radiation are the most important factors for the maintaining of the important functions of the existence of Biosphere. [2] It is well known that the solar radiation is the most important factor both for the process of photosynthesis of the plant world and of the actual temperature of the atmosphere. [2]

The presence of pollutants with the green house effect in the atmosphere leads to the increase of the global average atmospheric temperature. [3], [4] One of the largest components with the green house effect is carbon dioxide, which accounts for 75% from all pollutant gases. [4]

The solar radiation that interacts with the Earth's surface and atmosphere, a part of it is absorbed and another part is reflected. The reflected quantity has another conventional name as "albedo". [5], [6]

The albedo value of the oceans increases as the result of the decrease in the pH value of the ocean water. [7]

The increasing of pollutant green house gases leads to the increasing of the albedo of clouds and in its turn to the increasing of global albedo of Earth.[8]. The increasing of aerosol concentration leads to higher cloud droplet concentration, smaller cloud droplets, and higher cloud albedo.[9], [10].

The observatories of NASA register an increase in both carbon dioxide and albedo levels. It is observed that the albedo values increase over time. [11] The registrations of NASA are shown on the diagram of the Fig. 1 and the linear trend states in general that Earth's albedo is increasing with time.

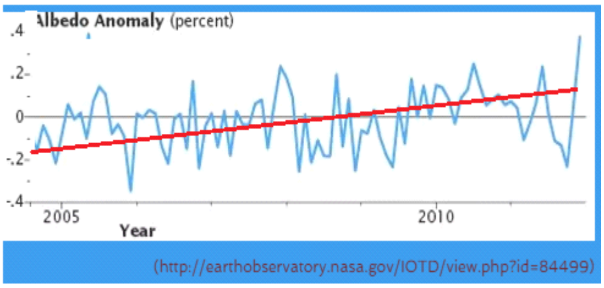

Figure 1: The increasing of Earth's albedo with time [11]

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The albedo values increase at the same time with the increasing of the green house effect. The processes of deforestation of the terrestrial surface lead to the increasing of the albedo value of the land. The average value of the albedo of the forest surfaces is of the order of 0.15. The other values for various terrestrial components are presented on the Table 1.

Table 1: The Albedo's values of various samples

Samples	Albedo values
Open ocean	0.06
Worn asphalt	0.12
Conifer forest(Summer)	0.08; 0.09 to 0.15
Deciduous forest	0.15 to 0.18
Bare soil	0.17
Green grass	0.25
Desert sand	0.40

[the results are based on the sources 12;13;14;15;16;17; 18].

The next Fig. 2 shows the maps of global deforestation. We can see from the maps of the Fig. 2 that the areas of deforestation are increasing with time. The places of bare soil have the values of albedo higher than of forests area. Another scientific paper [19] states that: "Starting with carbon dioxide concentration of 380 *ppm* and the increasing at the rate 0.2% /year within 100 years gives 456 ppm. The albedo will increase respectively from 0.3 till 0.36 within 100 years".

The current state of the atmosphere and of the biosphere in general presented in the above mentioned scientific works does not show any clear quantitative expression that links the albedo values and the pollutant concentrations of carbon dioxide in the atmosphere. Each result for both albedo and carbon dioxide and oxygen concentrations is presented separately and is described mostly only from a statistical-informational point of view. It is mentioned only that the temperature rises, but the reasons that lead to the temperature rise are not presented very clearly in the scientific papers.

The albedo values vary over time, but the reasons of these variations are also not clearly discussed. The processes that take place in the atmosphere are really complex and remain a complex study problem. But the most important aspect is that the current state of each parameter of the atmosphere is influenced by the values of other parameters. For example, how does the variation of atmospheric temperature influence albedo values? It is known that the increasing of the concentration of carbon dioxide leads to an increase in the average temperature of the atmosphere. In its turn, the logical question is: how does the increase in the concentration of carbon dioxide influence the albedo values? These major questions are the main tasks of the study of this current scientific research.

The interconnection between important parameters such as the values of the masses of carbon dioxide and oxygen in the atmosphere, the temperature of the atmosphere and the albedo values is a key issue of this paper.

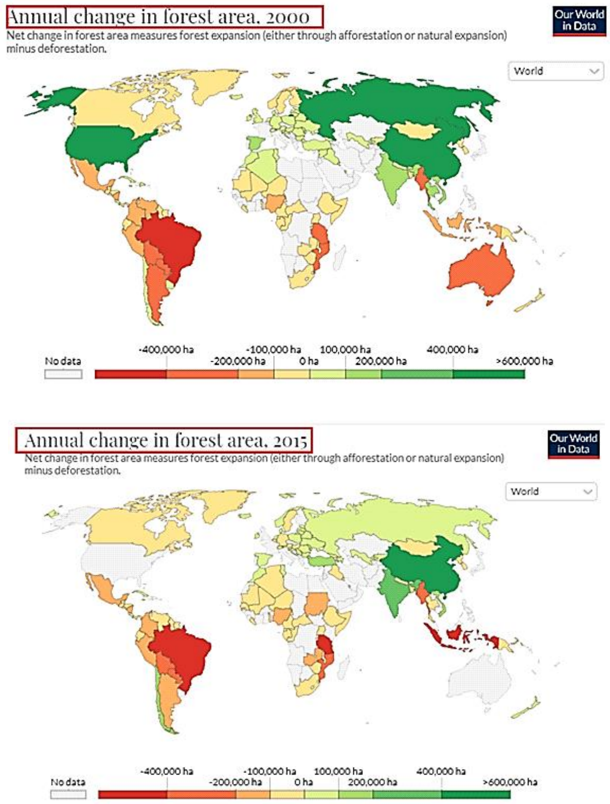

Figure 2: The maps of global deforestation [20]

2. Materials and methods

Each substance existing in terrestrial conditions has various values of albedo depending on chemical composition. [21] In general the sunlight carries energy to the planets of solar system. That sunlight is absorbed by the planet's surface heats the ground. [22] The planet will continue to warm until the outgoing infrared energy exactly balances the incoming energy from the sunlight. This balance is called "thermal equilibrium". [23] Satellites measure the amount of energy arriving at Earth from Sun as sunlight. This amount of incoming energy from sunlight is called "insolation". [23] This specific value of I_{sc}=1370 W/m^2 for Earth is called the "solar constant" and this constant is exactly for the case without air and water. In order to calculate the total amount of energy arriving at Earth is necessary to know how much area is being lit. This area multiplied by insolation gives the total amount of incoming energy. The amount of light intercepted (E_{interc}) by our spherical planet is exactly the same as the amount that would be blocked by a flat disk with the same diameter as Earth.[24]

$$E_{int\,erc} = I_{sc} \cdot \pi \cdot R_e^2; \quad R_e = 6371(km) \qquad (1)$$

$$E_{int\,erc} = 1370(W/m^2)\cdot 3.14\cdot(6.3171)^2.10^{12}(m^2) = 173.46\cdot 10^{15}(Watts) = 174(pentaWatts) \quad (2)$$

Since Earth is not completely black, then some of energy is reflected by the planet. This reflection is called namely “albedo”. In order to find how much energy the Earth absorbs from sunlight, then according to the law of energy: one minus albedo (A) equals to the light energy absorbed:

$$E_{absorbed} = I_{sc}\cdot(1-A)\cdot\pi\cdot R_e^2 \quad (3)$$

In order to calculate the energy flowing out from Earth is necessary to use the Stefan- Boltzman law that shows how much infrared energy is emitted from one unit of area:

$$E_{emitted} = \varepsilon\cdot\sigma\cdot T^4\cdot R_e^2 \quad (4)$$

ε-emissivity of Earth *(ε~0,6÷1)* depending on the type of the ground. The absolute black body has the value $\varepsilon=1$. The value $\varepsilon=1$ corresponds to the case of the vacuum (there is no atmosphere on Earth). The law of conservation of energy states that the energy emitted must be equal to the energy absorbed.

$$E_{emitted} = E_{absorbed} \quad (5)$$

$$I_{sc}\cdot(1-A) = 4\cdot\varepsilon\cdot\sigma\cdot T^4;\quad T=\sqrt[4]{\frac{I_{sc}(1-A)}{4\cdot\varepsilon\cdot\sigma}} \quad (6)$$

For the case when $\varepsilon=1$, then $T=T_{eff}$; T_{eff} - effective temperature; *A=0,3* for the Earth’s albedo:

$$T_{eff} = \sqrt[4]{\frac{I_{sc}(1-A)}{4\cdot\sigma}} = \sqrt[4]{\frac{1370\cdot(1-0.3)}{4\cdot 5{,}6704\cdot 10^{-8}}} \approx 255(K) \quad (7)$$

The effective temperature T_{eff} of Earth is such temperature when the Earth’s surface is considered as the black body with the condition that there is no any atmosphere around Earth.

Earth’s actual global average temperature is 15 ^{0}C (288 K). Our planet is warmer than predicted value of T_{eff} =255K by 33 ^{0}C. It is explained by the fact that certain gases in the atmosphere have green house effect leading to the increasing of the temperature till the value of 15 ^{0}C (288 K). The expression of the value of albedo *A* from (6) is:

$$A = 1-\frac{4\cdot\sigma\cdot\varepsilon}{I_{sc}}\cdot {T_{eff}}^4;\quad \varepsilon\approx 1 \quad (8)$$

The expression of the quantity of heat from Sun that is necessary for the atmosphere to maintain the value of actual average global temperature T_{atm} (T_{atm} *≈288 K)* is [4]:

$$Q = c_{air}.m_{air}.(T_{atm}-T_{eff});\quad \Rightarrow T_{eff} = T_{atm}-\frac{Q}{c_{air}.m_{air}} \quad (9)$$

where $m_{air}=5{,}148.10^{18}$ *kg* is the mass of Earth's atmospheric air; $c_{air}=1006$ *(J/(kg.K)* - specific thermal capacity of atmospheric air. [4] The substitution of the expression (8) into (9) gives the following:

$$(1-A)^{1/4} = \left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}\cdot T_{atm} - \left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}.\frac{Q}{c_{air}\cdot m_{atm}} \quad (10)$$

The expression (10) can be generalized and applied to the various substances with various thermal capacities *c* and masses. The value of density ρ can be applied for various substances.

$$(1-A)^{1/4} = \left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}\cdot T_{atm} - \left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}.\frac{Q}{c\cdot\rho\cdot V} \quad (11)$$

The components of Earth are various as soils, rocks, stones, lakes, seas, oceans, forests, desserts, sands, ices, etc. and all of them have various values of albedo, thermal capacity, density and volumes. It is expected that more dense substances have higher albedo, due to of the fact that the more compacted and dense structures have more possibility to “reflect” the sunlight. The values of densities are inverse proportional to the values of specific thermal capacities. For more densely substances are not necessary big quantities of energy to increase the temperature by one Kelvin [25]. It means that for more densely substances is necessary less quantity of energy than of the less densely substances. The general form of the dependence of specific thermal capacity as the function of the density of substances has such form as of the Fig. 3. The graphical dependence of $(1-A)^{1/4} = f\left(\frac{1}{c}\right)$has the general decreasing linear form. (Fig.4) The extrapolation to the ordinate axis gives the value $\left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}\cdot T_{atm}$ with the consequent calculation of the emission capacity ε of Earth. The intersection point with the ordinate axis means such case $c\to\infty$. This is the case of “absolute black body” for $c\to\infty$. [25] The value of albedo $A\to 0$ is the case of the absolute black body. Then $(1\text{-A})^{1/4}\to 1$ and de facto this intersection point is equal to one. $\left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}\cdot T_{atm} = 1 \quad\Rightarrow\quad \varepsilon = \frac{I_{sc}}{4\cdot\sigma\cdot(T_{atm})^4}$

So, the values $c\to\infty$; $A\to 0$; $\varepsilon\to 1$ corresponds to the case of "absolute black body".[25]

The temperature of the atmosphere for the case of absolute black body can be calculated:

$$T_{atm} = \sqrt[4]{\frac{I_{sc}}{4\cdot\sigma}} = \sqrt[4]{\frac{1370}{4\cdot 5.67\cdot 10^{-8}}} \approx 278(K)$$

The real value of emission capacity of Earth is $\varepsilon\approx 0.9$ and the average temperature of atmosphere is:

$$T_{atm} = \sqrt[4]{\frac{I_{sc}}{4\cdot\varepsilon\cdot\sigma}} = \sqrt[4]{\frac{1370}{4\cdot 0{,}9\cdot 5.67\cdot 10^{-8}}} \approx 286(K)$$

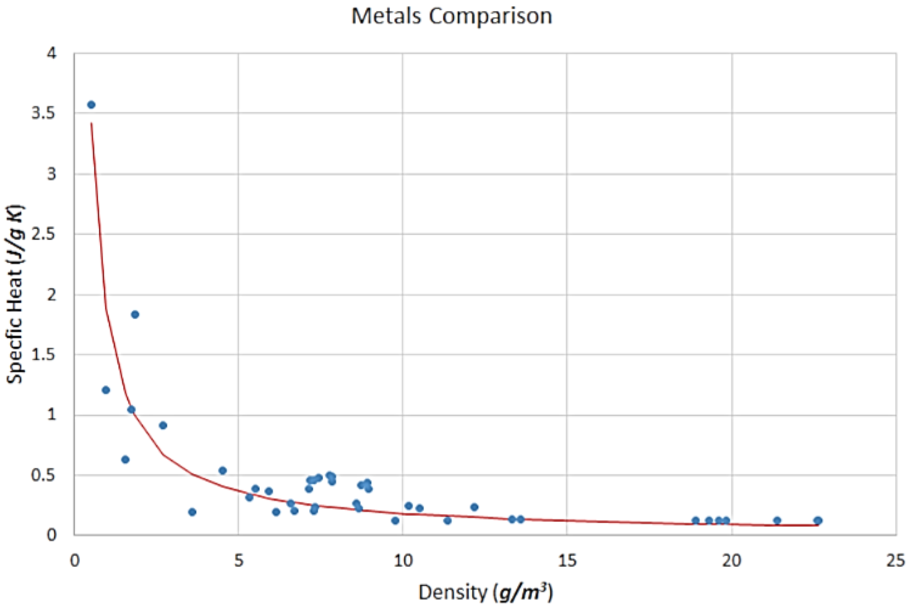

Figure 3: The general form of graphical dependence *c=f(ρ)* [26]

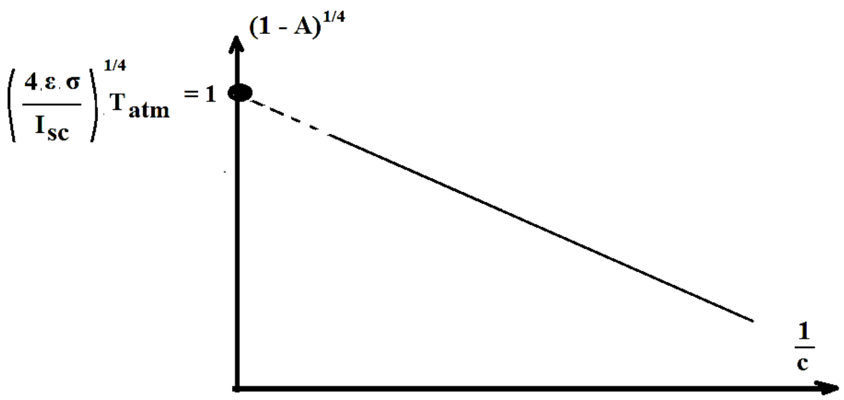

Figure 4: The general form of graphical dependence $(1-A)^{1/4} = f\left(\frac{1}{c}\right)$

3. Results and discussion

So, Earth's surfaces consist of various components with their values of densities, specific heat capacities and albedo. The data are presented on the Table 2.

Table 2: The constituents of Earth's surface with their values of densities, albedo and specific heat capacities

1	2	3	4	5	6	7	8	9	10	11
Components	**density; kg/m³**	**A**	**C; J/(kg/k)**	**1/C**	**T; K**	**Teff=T-Q/(cm)**	**(1-A) ¹/⁴**	**ΔT;K**	**lnC**	**ln(ΔT)**
vegetated land	1600	0.2	830	0.0012	294.8	272.22	0.94574	22.5832	6.72143	3.11721
wet sand (20% water)	2200	0.2	1500	0.00067	284.72	272.22	0.94574	12.496	7.31322	2.52541
average rock	2400	0.16	2000	0.0005	284.93	275.56	0.95735	9.37202	7.6009	2.23773
earth crust (thickness 30 km)	2200	0.3	1000	0.001	282.03	263.28	0.91469	18.744	6.90776	2.93088
forests	500	0.15	2500	0.0004	283.88	276.37	0.96018	7.49762	7.82405	2.01459
ice	917	0.7	2100	0.00048	221.95	213.02	0.88011	8.92573	7.64969	2.18894
desert sand	1900	0.4	980	0.00102	272.46	253.32	0.88011	19.1266	6.88755	2.95108
savannas, wet	700	0.2	900	0.00111	293.05	272.220	0.94574	20.8267	6.80239	3.03624
savannas, dry	1000	0.25	900	0.00111	288.69	267.86	0.9306	20.8267	6.80239	3.03624
tundra soils	250	0.2	900	0.00111	293.05	272.22	0.94574	20.8267	6.80239	3.03624
crops	750	0.25	700	0.00143	294.64	267.86	0.9306	26.7772	6.55108	3.28755

grassland	1340	0.25	800	0.00125	291.29	267.86	0.9306	23.4301	6.68461	3.15402
taiga soils	300	0.14	900	0.00111	298.01	277.18	0.963	20.8267	6.80239	3.03624
tropical rainforest	800	0.14	700	0.00143	303.96	277.18	0.963	26.7772	6.55108	3.28755
sea ice	916	0.6	1967	0.00051	238.44	228.90	0.8979	9.52925	7.58426	2.25437
steppe soils	1600	0.25	900	0.00111	288.69	267.86	0.9306	20.8267	6.80239	3.03624
clay	1400	0.23	920	0.00109	290.01	269.63	0.93675	20.374	6.82437	3.01426
Concrete, rough	1500	0.25	880	0.00114	289.16	267.86	0.9306	21.3	6.77992	3.05871
concrete, stone	1500	0.25	750	0.00133	292.86	267.86	0.9306	24.9921	6.62007	3.21856
dark soil	2000	0.1	900	0.00111	301.18	280.35	0.974	20.8267	6.80239	3.03624
Dark wet soil	1500	0.07	1480	0.00068	295.33	282.66	0.98202	12.6649	7.2998	2.53883
Dry clay	1000	0.23	1381	0.00072	283.2	269.63	0.93675	13.5728	7.23056	2.60807
dry earth	1400	0.22	1260	0.00079	285.38	270.50	0.93977	14.8762	7.13887	2.69976
dry sand	4000	0.35	840	0.00119	280.76	258.45	0.8979	22.3143	6.7334	3.10523
dry soil	1000	0.16	800	0.00125	298.99	275.56	0.95735	23.4301	6.68461	3.15402
granite	2700	0.22	790	0.00127	294.23	270.50	0.93977	23.7266	6.67203	3.1666
Graphite (carbon)	2500	0.12	717	0.00139	304.93	278.78	0.96855	26.1423	6.57508	3.26356
grey soil	1600	0.15	900	0.00111	297.2	276.37	0.96018	20.8267	6.80239	3.03624
lava	3100	0.19	840	0.00119	295.38	273.06	0.94868	22.3143	6.7334	3.10523
Light dry soil	1100	0.17	800	0.00125	298.17	274.73	0.95449	23.4301	6.68461	3.15402
Lime	3340	0.25	909	0.0011	288.48	267.86	0.9306	20.6205	6.81235	3.02629
Limestone	2750	0.25	909	0.0011	288.48	267.86	0.9306	20.6205	6.81235	3.02629
loam	1500	0.2	1600	0.00063	283.94	272.22	0.94574	11.715	7.37776	2.46087
peat lands	1400	0.16	1880	0.00053	285.53	275.56	0.95735	9.97023	7.53903	2.2996
quartz sand	1500	0.25	820	0.00122	290.72	267.86	0.9306	22.8586	6.7093	3.12933
Redland	1200	0.17	1040	0.00096	292.76	274.73	0.95449	18.0231	6.94698	2.89166
sand	1700	0.25	870	0.00115	289.41	267.86	0.9306	21.5449	6.76849	3.07014
Sand - dry	1500	0.3	900	0.00111	284.11	263.28	0.91469	20.8267	6.80239	3.03624
sand, clay, (dry)	1600	0.25	909	0.0011	288.48	267.86	0.9306	20.6205	6.81235	3.02629
sand, clay, (wetted)	2100	0.25	710	0.00141	294.26	267.86	0.9306	26.4001	6.56526	3.27337
Sandstone	2200	0.3	710	0.00141	289.68	263.28	0.91469	26.4001	6.56526	3.27337
sea water	1030	0.06	3850	0.00026	288.29	283.42	0.974	4.86858	8.25583	1.5828
Silver, highly polished	10500	0.8	235	0.00426	272.25	192.48	0.66874	79.7619	5.45959	4.37905
slaked lime	1350	0.25	840	0.00119	290.18	267.86	0.9306	22.3143	6.7334	3.10523
Soil - dark wet	2050	0.13	1480	0.00068	290.65	277.98	0.96578	12.6649	7.2998	2.53883
Stainless steel	7820	0.5	490	0.00204	280.3	242.04	0.8409	38.2531	6.19441	3.64423
stones	2500	0.25	1000	0.001	286.61	267.86	0.9306	18.744	6.90776	2.93088
wet sand	1922	0.25	2090	0.00048	276.83	267.86	0.9306	8.96844	7.64492	2.19371
wet soil	1500	0.11	1480	0.00068	292.24	279.57	0.97129	12.6649	7.2998	2.53883
water	1000	0.06	4180	0.00024	287.9	283.42	0.974	4.48422	8.33807	1.50056
soil minerals	2650	0.12	730	0.00137	304.46	278.78	0.96855	25.6768	6.59304	3.24559
soil organic matter	1300	0.09	1900	0.00053	291	281.13	0.9767	9.86528	7.54961	2.28902
calcite	2710	0.28	840	0.00119	287.46	265.14	0.92116	22.3143	6.7334	3.10523
iron	7900	0.5	452	0.00221	283.51	242.04	0.8409	41.4691	6.11368	3.72495
gravely earth	2050	0.17	1840	0.00054	284.92	274.73	0.95449	10.187	7.51752	2.32111
magnetite	5177	0.35	752	0.00133	283.38	258.45	0.8979	24.9256	6.62274	3.21589
rock	2560	0.15	879	0.00114	297.7	276.39	0.96018	21.3243	6.77878	3.05985
mountains	3000	0.3	900	0.00111	284.11	263.28	0.91469	20.8267	6.80239	3.03624
gypsum	1088	0.15	2110	0.00047	285.26	276.37	0.96018	8.88343	7.65444	2.18419
asphalt	2322	0.17	960	0.00104	294.26	274.73	0.95449	19.525	6.86693	2.9717
Asphalt(compacted)	2360	0.16	900	0.00111	296.39	275.56	0.95735	20.8267	6.80239	3.03624
basalt rock	2700	0.16	840	0.00119	297.88	275.56	0.95449	22.3143	6.7334	3.10523
betone	1800	0.3	840	0.00119	285.6	263.28	0.91469	22.3143	6.7334	3.10523
black earth (chernozem)	1100	0.05	1800	0.00056	294.58	284.17	0.98726	10.4134	7.49554	2.34309

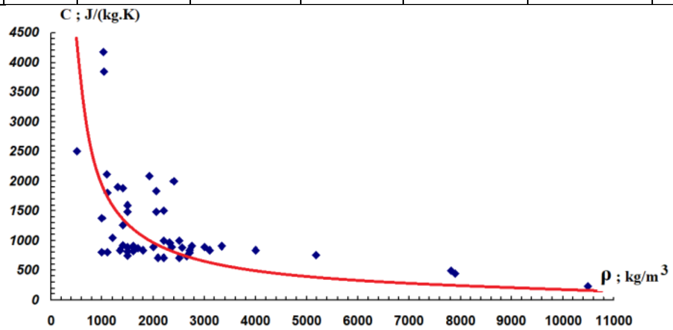

Figure 5: The dependence of the specific heat capacity as the function of densities of Earth's substances

The representation of the graphic of specific heat capacity as the function of densities of the Earth's components from the Table 1 is represented on the Fig.5. The shape of this dependence from Fig.5 has the similar dependence as Fig. 3. The values of Earth's albedo which are represented on the Table 1 suggest for the dependence of these albedo values of Earth's components as the function of the densities. (Fig. 6)

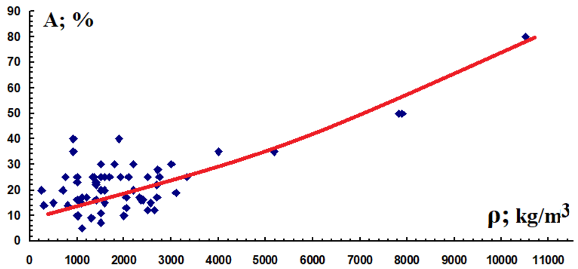

Figure 6: The dependence of the values of Albedo as the function of the densities

This situation suggests about the same similar dependence as Fig. 3 of the decreasing of the albedo values with the increasing of the values of specific heat capacities. (Fig. 7)

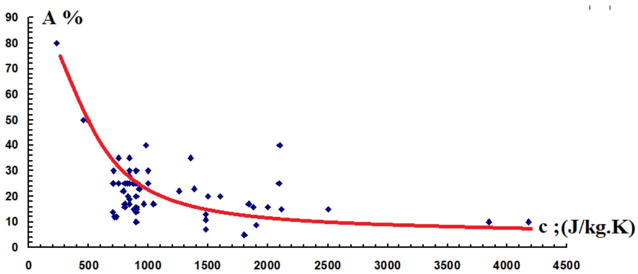

Figure 7: The dependence of the values of Albedo as the function of specific heat capacities

The presentation of the dependence of $(1-A)^{1/4} = f\left(\frac{1}{c}\right)$ has the form (Fig. 8):

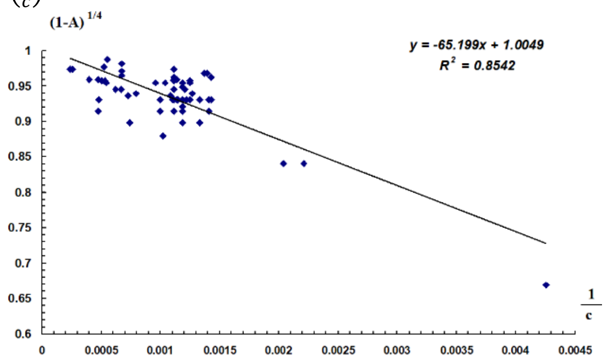

Figure 8: The dependence of $(1-A)^{1/4} = f\left(\frac{1}{c}\right)$

The correlational calculations of the graphic $(1-A)^{1/4} = f\left(\frac{1}{c}\right)$ shows that $\left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}\cdot T_{atm} \cong 1 \quad\Rightarrow\quad \varepsilon = \frac{I_{sc}}{4\cdot\sigma\cdot(T_{atm})^4}$; $T_{atm} = 288(K)$; $\varepsilon = \frac{I_{sc}}{4\cdot\sigma\cdot(T_{atm})^4} = \frac{1370}{4\cdot 5.67\cdot 10^{-8}\cdot(288)^4} \approx 0.88$

The case when the heat sunlight is absent (*Q=0*), then:

$$\left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}\cdot T = (1-0.3)^{1/4}$$

$$\boldsymbol{T^4 = \frac{0.7\cdot 1370}{4\cdot 0.88\cdot 5.67\cdot 10^{-8}} = 48.05\cdot 10^8 \quad\Rightarrow\quad T = \sqrt[4]{48.05\cdot 10^8} = 263(K)}$$

The special case ε→1 of black body, then T=255 (K).

Another coefficient of the linear dependence $(1-A)^{1/4} = f\left(\frac{1}{c}\right)$ shows that the correlational calculation is:

$$\left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}.\frac{Q}{\rho\cdot V} = 65.199 \Rightarrow \frac{Q}{m} = \frac{65.199\cdot(I_{sc})^{1/4}}{(4\cdot\varepsilon\cdot\sigma)^{1/4}} = 65.199\cdot\sqrt[4]{\frac{1370}{4\cdot 0.88\cdot 5.67\cdot 10^{-8}}} = 18744.04(J/kg)$$

In order to validate this result $\frac{Q}{m} = 18744.04(J/kg)$ is necessary to use this value for checking of the known value such as the average temperature of the Earth's atmosphere T_{atm}=288 (K). It means that the application of this value 18744.04 (J/kg) must give the similar value of the temperature of the order 288 (K). The application of the expression (11) gives the following expression:

$$T_{atm} = \frac{Q}{m.c} + \sqrt[4]{\frac{(1-A)\cdot I_{sc}}{4\cdot\varepsilon\cdot\sigma}}; \quad \frac{Q}{m} = 18744.04(J/kg); \quad \varepsilon = 0.88;$$

The calculated values of T_{atm} are represented on the sixth column of the Table 2.

The statistical χ² - test gives the answer about the normal (Gauss) distributed values of the temperatures. The performed calculations about this test are represented on the table 3.

The value $\chi^2 = \sum\frac{(E-O)^2}{E}$. If $\chi^2 \le \chi^2_{cr}$ then the sample of data is distributed by Gauss distribution. The critical value of χ^2_{cr} is performed by the statistical soft and this value is $\chi^2_{cr} = 12{,}5916 \quad\Rightarrow\quad \chi^2 < \chi^2_{cr}$

This result shows that the values of temperatures are distributed by normal distribution. The respective statistical histogram of Gauss distribution is represented on the Fig. 9

The maximum of the distribution corresponds to the average value of the temperature *<T>=289.13 (K).*

The expression (11) : $(1-A)^{1/4} = \left(\frac{4.\varepsilon.\sigma}{I_{sc}}\right)^{1/4}\cdot\left(T_{atm} - \frac{Q}{c\cdot m}\right) = \left(\frac{4.\varepsilon\sigma}{I_{sc}}\right)^{1/4}\cdot T_{eff}$ allows to represent the dependence $(1-A)^{1/4} = f(T_{eff})$. The values of effective temperatures T_{eff} are calculated if the value $\frac{Q}{m} = 18744.04(J/kg)$ is known and the values of specific heat capacities *c* are taken from the Table 1 for various Earth's components.

Table 3: The χ^2 - test of the normality of the sample

cells	cell start	cell end	cell interval	probability	E-expected frequency	O-observed frequency	$(E-O)^2/E$
1	262.255	268.351	[262---268]	0.004161	0.266321	0	0.266320
2	268.351	274.447	[268---274]	0.028212	1.805591	3	0.790108
3	274.447	280.543	[274---281]	0.108695	6.956494	2	3.531496
4	280.543	286.639	[281---287]	0.23844	15.26015	19	0.916534
5	286.639	292.735	[287---293]	0.298184	19.08379	16	0.498315
6	292.735	298.831	[293---299]	0.212661	13.61033	18	1.415777
7	298.831	304.927	[299---305]	0.086447	5.532595	5	0.051270
							χ^2=7,469

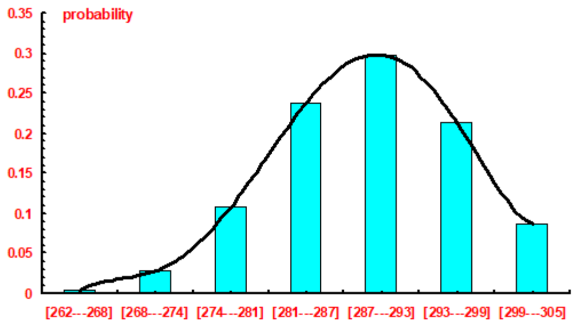

Figure 9: The distribution of Gauss of the temperature values

The graphic of the dependence $(1-A)^{1/4} = f(T_{eff})$ is represented as:

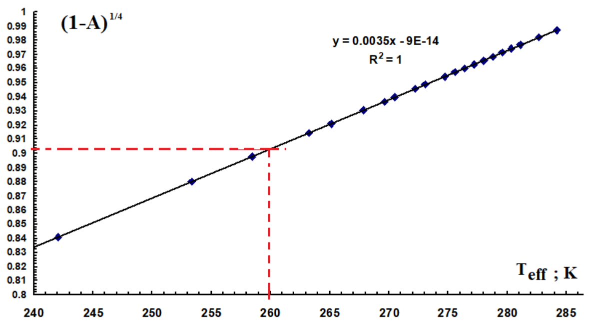

Figure 10: The graphic of the dependence $(1-A)^{1/4} = f(T_{eff})$

If the average value of Earth albedo is *A=0.33*, then $(1-A)^{1/4} = 0.904729 \quad \Rightarrow T_{eff} \approx 260(K)$

The proportional coefficient of the correlational dependence is 0.0035. This value of this proportional coefficient is $\left(\frac{4.\varepsilon\sigma}{I_{sc}}\right)^{1/4} = 0.0035 \quad \Rightarrow \varepsilon = \frac{1370\cdot(0.0035)^4}{4\cdot5.67\cdot10^{-8}} \approx 0.9$

The effective temperature of Earth officially is 255 K [27]. The obtained result is ≈260 K. The value of 255 (K) is calculated for the case of absolute "black body" with the value of emmisivity *ε=1.* The calculated result is *ε=0.9,* therefore the effective temperature is a little bit higher than for the case of *ε=1.* The graphic of the dependence of $T_{eff}=f(A)$ is represented on Fig. 11. As we can see the Figure 11 we can conclude that as higher is the albedo value then smaller is the effective temperature. This situation is described in [28] as: "All else being equal, temperatures will be warmer over a low albedo surface than over a high albedo surface." The representation of the values of the temperatures both for the effective values and equilibrium temperatures *T* are represented on the same plot of the Fig. 12.

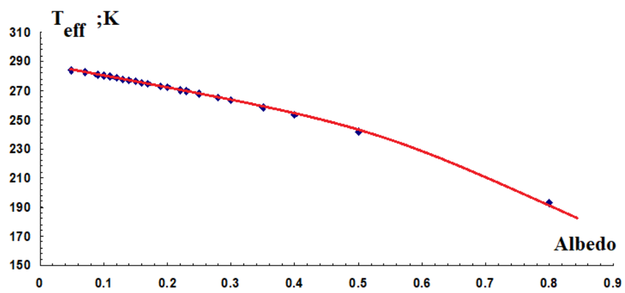

Figure 11: The values of effective temperatures of Earth's components as the function of albedo values

It can be observed that higher albedo values give larger interval between equilibrium value of the temperature and effective one. It means that components with higher albedo are cooling more rapidly for the case of the absence of sunlight. Therefore, the desert sands with relative higher albedo values have extreme variations of the temperatures from minimal to maximal values of the order of differences as 30 ^{0}C. [28, 29] This difference can be seen on the Fig. 12.

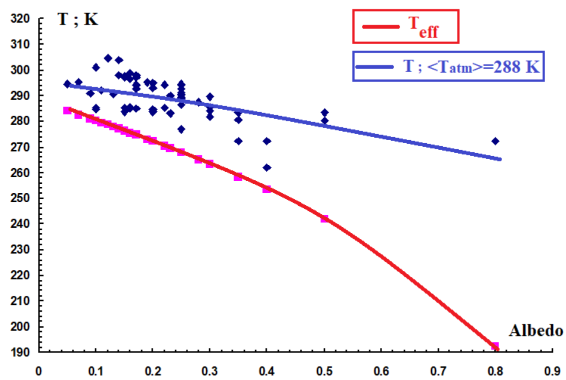

Figure 12: The values of effective and equilibrium temperatures of Earth's components as the function of albedo values

The difference of temperatures $\Delta T=T-T_{eff}$ as the function of albedo values are represented on Fig. 13.

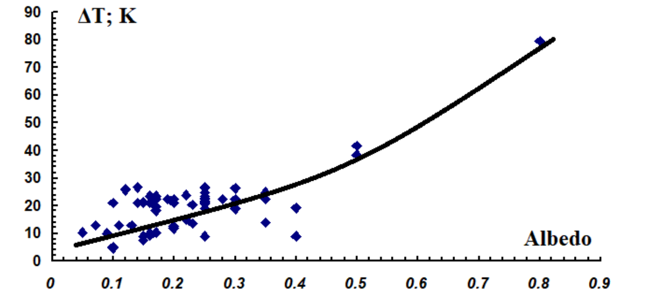

Figure 13: The values $\Delta T=T-T_{eff}$ as the function of albedo values

The values $\Delta T=T-T_{eff}$ as the dependence of specific heat capacities is represented on Fig. 14. Regarding the Figure 14, it can be observed that the differences of temperatures ΔT are smaller for the higher values of specific heat capacities and respectively for smaller values of albedo.

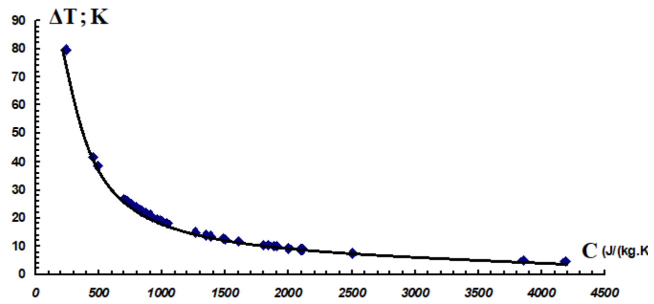

Figure 14: The values $\Delta T=T-T_{eff}$ as the function of specific heat capacities

The expression of heat Q as the function of the temperature variations ΔT is [4]:

$$\boldsymbol{Q = c.m.\Delta T \quad \Rightarrow \quad \ln \Delta T = \ln\left(\frac{Q}{m}\right) - \ln C}$$

The graphic of the dependence of $\ln \Delta T = f(\ln C)$ is represented on Figure 15.

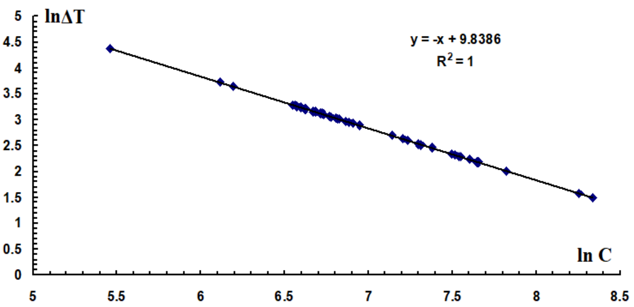

Figure 15: The graphic of the dependence $\ln \Delta T = f(\ln C)$

The graphic of the dependence $\ln \Delta T = f(\ln C)$ allows to find the value $\frac{Q}{m}$. The correlational calculations of this dependence shows that:

$$\boldsymbol{\ln\left(\frac{Q}{m}\right) = 9.8386 \quad \Rightarrow \quad \frac{Q}{m} = e^{9.8386} = 18743.46\left(\frac{J}{kg}\right)}$$

The validity of the value $\frac{Q}{m} = 18744\left(\frac{J}{kg}\right)$ can be checked by the real case of the surface of **Earth:**

$$T = \frac{Q}{m.c} + \sqrt[4]{\frac{(1-A).I_{SC}}{4.\varepsilon\sigma}} = \frac{18744\left(\frac{J}{kg}\right)}{543\left(\frac{J}{kg.K}\right)} + \sqrt[4]{\frac{(1-0.32)\cdot 1370}{4\cdot 0.9\cdot 5.67\cdot 10^{-8}}} \approx 294(K) \equiv 21^0C$$

The specific heat capacity of planet Earth is 543 (J/kg.K) [29]. The actual Earth's surface average temperature is 288 K. The respective relative error is

$$\varepsilon\% = \frac{|294-288|}{291}\cdot 100\% \cong 2\%$$

This calculated value $\frac{Q}{m} = 18744\left(\frac{J}{kg}\right)$ could be called as the specific Earth's heat content. It shows the quantity of Sun heat absorbed by 1 kg of substance in order to maintain the difference of temperatures $\Delta T=T-T_{eff}$.

Another calculation is the Ocean's water:

$$T = \frac{Q}{m.c} + \sqrt[4]{\frac{(1-A).I_{SC}}{4.\varepsilon\sigma}} = \frac{18744\left(\frac{J}{kg}\right)}{3850\left(\frac{J}{kg.K}\right)} + \sqrt[4]{\frac{(1-0.06).1370}{4\cdot 0.85\cdot 5.67\cdot 10^{-8}}} \approx 291(K) \equiv 18^0C$$

The specific heat capacity of seawater is 3850 (J/kg.K) [30]. The emissivity of seawater is ε=0.85 [31]. The obtained result of the temperature is of order of real 17^{0}C of sea water.[32]

- **Sea ices:** Albedo of sea ices is 0.6 [33]; emisivity of sea ices is ε=0.9 [34]; c=1967 (J/(kg.K))

$$T = \frac{Q}{m.c} + \sqrt[4]{\frac{(1-A).I_{SC}}{4.\varepsilon\sigma}} = \frac{18744\left(\frac{J}{kg}\right)}{1967\left(\frac{J}{kg.K}\right)} + \sqrt[4]{\frac{(1-0.6).1370}{4\cdot 0.9\cdot 5.67.10^{-8}}} \approx 237(K) \equiv -36^0C$$

- **Air:** The emisivity of air is $\varepsilon=0.8$ [35]. The albedo of air $A=0.3$; thermal capacity of air c =1007 J/(kg.K)

$$T = \frac{Q}{m.c} + \sqrt[4]{\frac{(1-A).I_{SC}}{4.\varepsilon\sigma}} = \frac{18744\left(\frac{J}{kg}\right)}{1007\left(\frac{J}{kg.K}\right)} + \sqrt[4]{\frac{(1-0.3)\cdot 1370}{4\cdot 0.8\cdot 5.67\cdot 10^{-8}}} \approx 288(K) \equiv 15^0C$$

The recalculations of the temperatures of other Earth's components are represented on the Table 4.

Table 4: The calculation of the temperatures of Earth's components

Earth's components	Emissivity - ε	C; (J/(kg.K))	Albedo	T_{eff}; K	T; K	ΔT; K	lnΔT	lnC
Earth	0.9	543	0.32	259.917	294.437	34.519	3.5415	6.2971
air	0.8	1007	0.3	269.631	288.245	18.613	2.9238	6.9147
Sea ices	0.9	1967	0.6	227.626	237.156	9.5292	2.2543	7.5842
Ocean's water	0.85	3850	0.06	285.888	290.757	4.8685	1.5828	8.2558
desert sand	0.9	980	0.4	251.910	271.037	19.126	2.9510	6.8875
mountains	0.96	900	0.3	257.617	278.444	20.826	3.0362	6.8023
woodland	0.985	1360	0.16	267.904	281.686	13.782	2.6233	7.2152
grassland	0.982	900	0.16	268.108	288.935	20.826	3.0362	6.8023
arable land	0.975	1100	0.1	273.261	290.301	17.04	2.8355	7.0030
bare land	0.972	1000	0.2	265.537	284.281	18.744	2.9308	6.9077
dry grassland	0.91	1000	0.15	274.071	292.815	18.744	2.9308	6.9075
fresh snow	0.9	2090	0.8	191.410	200.379	8.9684	2.1937	7.6449
dark soil, high humus	0.9	1100	0.1	278.784	295.824	17.04	2.8355	7.0030

The data from the Table 4 serve to represent the following dependence $ln(\Delta T) = f(ln(C))$ on the Figure 16.

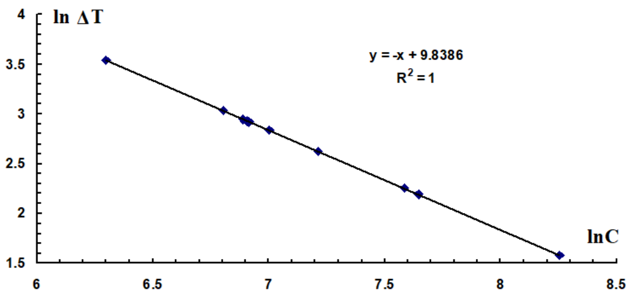

Figure 16: The graphic of the dependence $ln\,\Delta\,T = f(ln\,C)$ based on the data of the Table 5

The correlational calculations of this dependence shows again that the value of specific Earth's heat content is: $\frac{Q}{m} = \mathbf{18743.46}\left(\frac{J}{kg}\right)$**.** In order to obtain the expression of the variation of Albedo as the dependence of the variation of the temperature of the atmosphere *ΔTa* by the influence of pollutants such as carbon dioxide and the variation of emissive thermal energy - ΔE, it is necessary to perform the derivative operation (*A/*) (differentiation (*dA*)) of the expression of Albedo:

$$A = 1 - \frac{4 \cdot \varepsilon \cdot \sigma}{I_{sc}} \cdot \left(T_a - \frac{Q}{m.c}\right)^4$$

$$A^{-} = dA = \Delta A = -\frac{16 \cdot \varepsilon \cdot \sigma}{I_{sc}} \cdot T_{eff}^3 \cdot \left(\Delta T_a - \frac{1}{c} \cdot \left(\frac{\Delta Q}{m} - \frac{Q \cdot \Delta m}{m^2}\right)\right) \Delta A = \frac{16 \cdot \varepsilon \cdot \sigma}{I_{sc}} \cdot T_{eff}^3 \cdot \left(\frac{1}{m \cdot c} \cdot \left(\Delta Q - \frac{Q \cdot \Delta m}{m}\right) - \Delta T_a\right)$$

The variation of the mass Δm is related to the resultant change due to of the depletion of oxygen in the atmosphere $\Delta m(O_2)$ and the accumulation of carbon dioxide $\Delta m(CO_2)$. [4] Their rates are presented on the Table 5. The value *ΔTa* is the increasing of the temperature of atmospheric air due to of the presence of excess antropogenic pollutant green house gas - carbon dioxide. [4] This value is of the order ~ 0.025 °C/year. [4] The resultant accumulated heat ΔQ due to of the main complex processes in the atmosphere as: 1) anthropogenical burning of fuels; 2) respiration of Land biota together with the human population; 3) photosynthesis. This value is of the order $\Delta Q = 3.1 \cdot 10^{20}$ Joule/year[4]. Only 40% of this quantity of heat is absorbed by the atmosphere, but another 60% is absorbed by Oceans. [4] Taking into consideration the mass of atmosphere $m = 5.13 \cdot 10^{18}$ kg and the value of specific Earth's heat content $\frac{Q}{m} = 18744.04(J/kg)$, then the calculated values of the variation of albedo are represented on the Table 5.

The values ΔT_{eff} from the Table 5 are the differences of effective temperatures between actual fixed value and of each consequent value of effective temperature. The value of long wave thermal radiation E [w/m^2] is calculated by the Stefan-Boltzmann law : $E = \sigma \cdot T_{eff}^4$

The dependence of the variation of albedo ΔA of the system atmosphere-Earth as the function of the variation of the global average atmospheric temperature ΔT_a is represented on the Figure 17. the rising of the atmospheric temperature leads to the increasing of the albedo of the system atmosphere-Earth. The rate of rising of albedo is 10^{-5}/ ^{0}C. (Figure 17). The effective temperature T_{eff} of this complex system atmosphere-Earth respectively is decreasing by module with the respective increasing of the global average atmospheric temperature ΔT_a. (Figure 18). The rate of decreasing of the effective temperature of this complex system atmosphere-Earth is ~ 0.0012 ^{0}C for each 1^{0}C rising of the atmospheric temperature.

Table 5: The dynamics of change of Albedo values and effective temperature with time (columns 1-5 are based on the data from the source [4])

1	2	3	4	5	6	7	8	9	10
year	ΔT; oC	ΔQ; Joule	Δm(CO2); kg	depletionΔ m(O2); kg	ΔA	A	T_{eff};K	ΔT_{eff};K	E;[w/m^2]
1981	0.03	1.22E+20	1.081E+13	1.83E+13	-2.0803E-07	0.2999998	255.0022	1.314E-05	239.75007
1982	0.05	2.44E+20	1.621E+13	1.85E+13	-6.3728E-08	0.2999999	255.00218	3.673E-05	239.75002
1983	0.07	3.66E+20	2.701E+13	2E+13	1.95303E-07	0.3000002	255.00216	5.913E-05	239.74993
1984	0.09	4.88E+20	3.782E+13	2.2E+13	4.41231E-07	0.3000004	255.00214	8.414E-05	239.74985
1985	0.11	6.1E+20	4.862E+13	2.3E+13	7.15897E-07	0.3000007	255.00211	9.751E-05	239.74975
1986	0.13	7.32E+20	5.403E+13	2.32E+13	8.62654E-07	0.3000009	255.0021	0.0001099	239.7497
1987	0.15	8.54E+20	5.943E+13	2.38E+13	9.98671E-07	0.300001	255.00209	0.0001228	239.74966
1988	0.18	9.76E+20	6.483E+13	2.42E+13	1.14074E-06	0.3000011	255.00207	0.0001289	239.74961
1989	0.19	1.098E+21	6.753E+13	2.46E+13	1.2073E-06	0.3000012	255.00207	0.000135	239.74959
1990	0.22	1.22E+21	7.023E+13	2.5E+13	1.27407E-06	0.3000013	255.00206	0.000148	239.74956
1991	0.26	1.342E+21	7.564E+13	2.54E+13	1.41726E-06	0.3000014	255.00205	0.0001611	239.74951
1992	0.32	1.464E+21	8.104E+13	2.58E+13	1.5609E-06	0.3000016	255.00204	0.0001879	239.74947
1993	0.34	1.586E+21	9.184E+13	2.63E+13	1.85507E-06	0.3000019	255.00201	0.0002148	239.74936
1994	0.36	1.708E+21	1.026E+14	2.68E+13	2.15017E-06	0.3000022	255.00198	0.0002425	239.74926
1995	0.37	1.83E+21	1.135E+14	2.7E+13	2.45473E-06	0.3000025	255.00195	0.0002561	239.74916
1996	0.39	1.952E+21	1.189E+14	2.73E+13	2.60377E-06	0.3000026	255.00194	0.0002681	239.74911
1997	0.41	2.074E+21	1.243E+14	2.82E+13	2.73618E-06	0.3000027	255.00193	0.0002795	239.74906
1998	0.42	2.196E+21	1.297E+14	2.94E+13	2.86045E-06	0.3000029	255.00192	0.0002929	239.74902
1999	0.43	2.318E+21	1.351E+14	2.98E+13	3.00795E-06	0.300003	255.0019	0.000321	239.74897
2000	0.45	2.44E+21	1.459E+14	3E+13	3.31625E-06	0.3000033	255.00188	0.0003752	239.74886
2001	0.46	2.562E+21	1.675E+14	3.1E+13	3.9123E-06	0.3000039	255.00182	0.0004297	239.74866
2002	0.47	2.684E+21	1.891E+14	3.2E+13	4.51021E-06	0.3000045	255.00177	0.0004552	239.74846
2003	0.5	2.806E+21	1.999E+14	3.33E+13	4.79067E-06	0.3000048	255.00174	0.0004693	239.74836
2004	0.54	2.928E+21	2.053E+14	3.36E+13	4.94521E-06	0.3000049	255.00173	0.0004974	239.74831
2005	0.58	3.05E+21	2.161E+14	3.4E+13	5.25315E-06	0.3000053	255.0017	0.0005108	239.7482
2006	0.59	3.172E+21	2.215E+14	3.46E+13	5.40018E-06	0.3000054	255.00169	0.0005242	239.74815
2007	0.6	3.294E+21	2.269E+14	3.52E+13	5.54765E-06	0.3000055	255.00167	0.0005371	239.7481
2008	0.61	3.416E+21	2.323E+14	3.6E+13	5.68976E-06	0.3000057	255.00166	0.0005488	239.74805
2009	0.62	3.538E+21	2.377E+14	3.73E+13	5.81777E-06	0.3000058	255.00165	0.0005621	239.74801
2010	0.63	3.66E+21	2.431E+14	3.8E+13	5.9636E-06	0.300006	255.00163	0.0005905	239.74796
2011	0.65	3.782E+21	2.539E+14	3.84E+13	6.2759E-06	0.3000063	255.00161	0.0007057	239.74785
2012	0.68	3.904E+21	2.971E+14	3.86E+13	7.54027E-06	0.3000075	255.00149	0.000778	239.74742
2013	0.7	4.026E+21	3.242E+14	3.88E+13	8.33514E-06	0.3000083	255.00142	0.0008219	239.74715
2014	0.74	4.148E+21	3.404E+14	3.9E+13	8.81637E-06	0.3000088	255.00137	0.000837	239.74698
2015	0.78	4.27E+21	3.458E+14	3.92E+13	8.98249E-06	0.300009	255.00136	0.000895	239.74692
2016	0.8	4.392E+21	3.674E+14	3.96E+13	9.61879E-06	0.3000096	255.0013	0.0009515	239.74671
2017	0.81	4.514E+21	3.89E+14	4.06E+13	1.02394E-05	0.3000102	255.00125	0.0009808	239.74649
2018	0.82	4.636E+21	3.998E+14	4.1E+13	1.05615E-05	0.3000106	255.00122	0.0009952	239.74638
2019	0.83	4.758E+21	4.052E+14	4.16E+13	1.07193E-05	0.3000107	255.0012	0.0010247	239.74633
2020	0.85	4.88E+21	4.16E+14	4.2E+13	1.10431E-05	0.300011	255.00117		239.74622

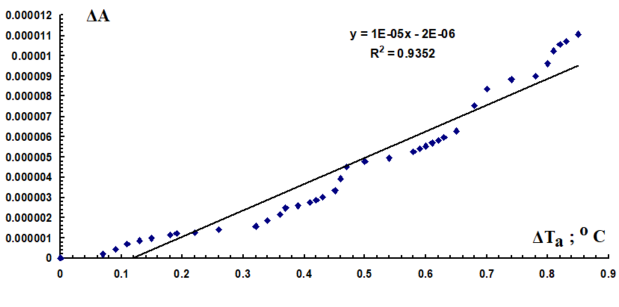

Figure 17: The graphic of the dependence $\Delta A = f(\Delta T_a)$

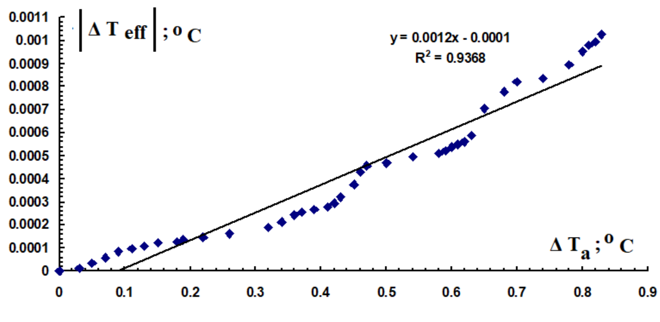

Figure 18: The graphic of the dependence $|\Delta T_{eff}| = f(\Delta T_a)$

The outgoing energy of wavelength radiation from atmosphere-Earth system that is described by the value E[W/m^2) shows that are declining with the increasing of the albedo values of this complex system. (Figure 19). The correlational calculations of the Figure 19 show that the proportional coefficient of the linear dependence is 342.5. If the value *A* which is the independent variable *x* of this linear dependence is zero (*x=0*) then *E=342.5 (W/m^2)* and this case is exactly the case of "absolute black body". The effective temperature of this case of "absolute black body" is calculated from the Stefan-Boltzmann law:

$$T_{eff} = \sqrt[4]{\frac{342.5}{5.67 \cdot 10^{-8}}} = 278.78(K)$$

The case when *y=0* then albedo *A=1* and this is the case of absolute reflectance of energy.

It can be seen from Figure 19 that if the value of Albedo is increased by 0.01 then *E* decreases by 3.42 W/m^2

The same order of the result 3.4 W/m^2 is presented by studies [36], [37] with such statements: "the increasing of albedo by 0.01 leads to the decreasing of energy by 3.4 W/m^2 ". The energy of outgoing wavelength radiation is of order 235 w/m^2 and is represented in the study [38] (Figure 20).

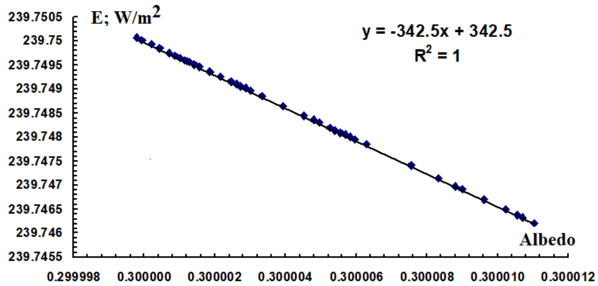

Figure 19: The graphic of the dependence *E=f(A)*

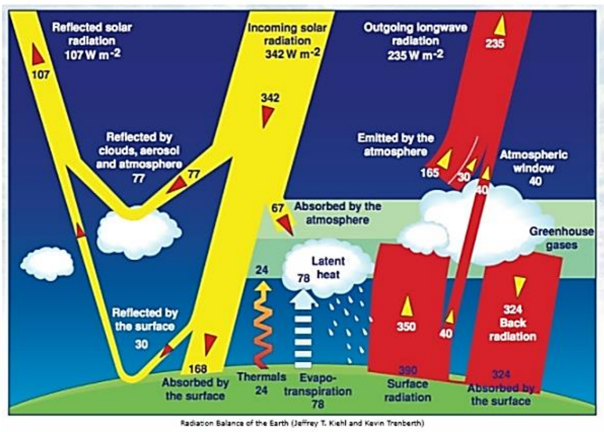

Figure 20: The radiation balance of the Earth [38]

The graphic of the dependence of the values of effective temperatures as the function of albedo is represented on the Figure 21.

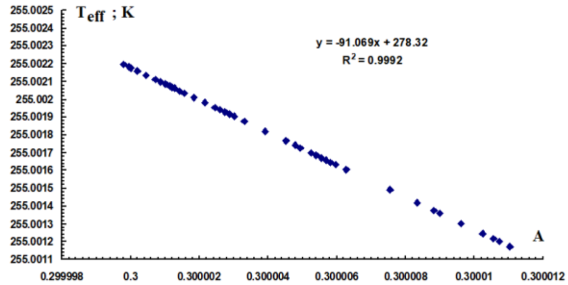

Figure 21: The dynamics of the evolution of effective temperatures with albedo values of Earth-atmosphere complex system

The analysis of Figure 21 shows that each increasing of albedo by 0.01, then the temperature decreases by 0,91 ^{o}C wich is of the order 1 ^{o}C that is presented in the paper [37] as: "it is estimated that the changing of albedo by 0.01 leads to the change of the temperature by 1 ^{0}C". The similar result of the decreasing of the level of outgoing longwave radiation is represented on the Figure 22. The application of Wien's law of displacement allows to calculate the values of wavelengths and to represent the graphic of the dependence *λ=f(A)*. (Figure 23)

The values of long wavelengths radiation are of the order 11 µm that are coincident with the values from paper [38]. The correlational calculations of the dependence λ=*f(A)* shows that $\lambda = 4.0574 \cdot A + 10.143$. Taking the Earth's albedo *A=0.3* then *λ=11.36* µm.

The decreasing of the effective temperatures with the increasing of the average atmospheric temperature is explained by the mechanism from the paper [39]. There is an important remark of the paper [39]: "If the amount of greenhouse gases were to increase without allowing the temperature of the surface-troposphere system to change, the upward flux of longwave radiation at the top of the atmosphere would decrease".

Regarding this remark, then the troposphere together with Earth's surface forms this complex system that is described by Albedo of this complex system and the effective temperature in fact is the temperature not directly of Earth's surface but of this system atmosphere-Earth. The daily registration of outgoing longwave radiation by satellites is represented on the Figure 24.

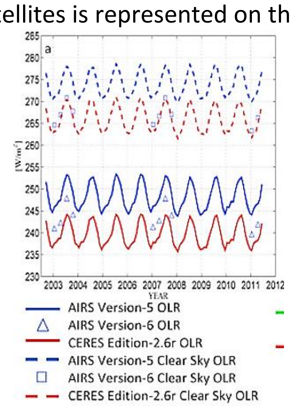

Figure 22: The evolution of Earth's longwave emission [40]

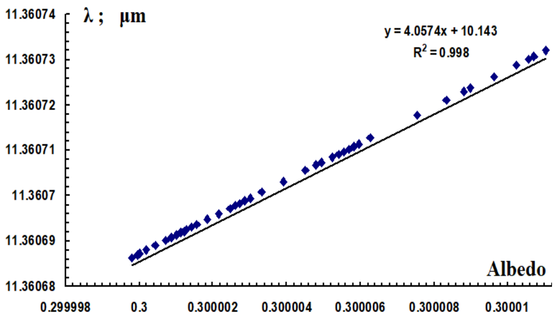

Figure 23: The dependence of Earth's wavelengths as the function of Albedo

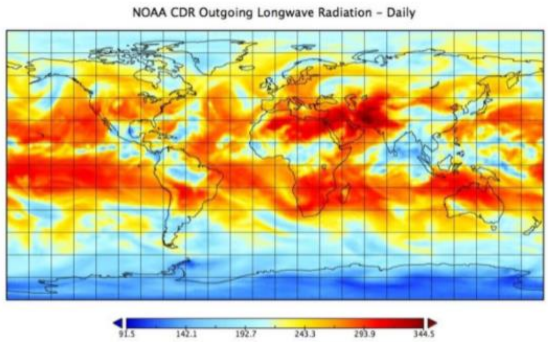

Figure 24: The map of registration of outgoing long wave radiation [41]

The distribution by geographical area of outgoing long wave radiation is represented on the Table 6. the average outgoing longwave radiation <OLR>= $\frac{\sum_i (OLR)_i \cdot S_i}{100\%}$; where S_i is the percent of area for the respective value of geographical place.

$$< OLR >= \frac{\sum_i (OLR)_i \cdot S_i}{100\%} = \frac{340 \cdot 15\% + 300 \cdot 15\% + 270 \cdot 25\% + 190 \cdot 30\% + 140 \cdot 15\%}{100\%} \approx 240(W/m^2)$$

and the obtained result is of order of the value from the Figure 19.

The interconnection between the wavelengths and the albedo values can be seen on the Figure 25.

Table 6: The distribution of outgoing lonwave radiation by geographical area

OLR; W/m²	340	300	270	190	140
% of surface	15	15	25	30	15

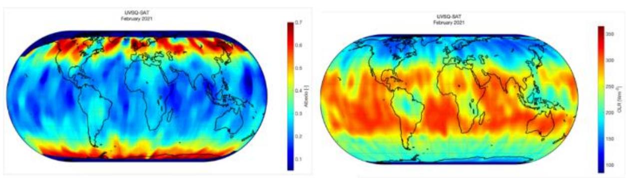

Figure 25: The interconnection of outgoing longwave radiation and albedo values [42]

The extracted values of albedo and energy (W/m²) are represented on the Table 7.

Table 7: The outgoing longwave radiation of Earth

A	E; W/m²	T_{eff}; K	E_{th}; W/m²	λ; μm
0.2	280	270.696	304.444	10.70205668
0.6	160	227.627	152.222	12.72696194
0.3	270	261.808	266.389	11.06535199
0.7	110	211.831	114.167	13.67601063
0.35	270	257.002	247.361	11.27227056
0.15	300	274.83	323.472	10.54107763
0.7	120	211.831	114.167	13.67601063
0.25	230	266.363	285.417	10.87613094

The values of effective temperatures are calculated (Table 7). The dependence of effective temperatures T_{eff} as the function of albedo values is represented on the Figure 26.

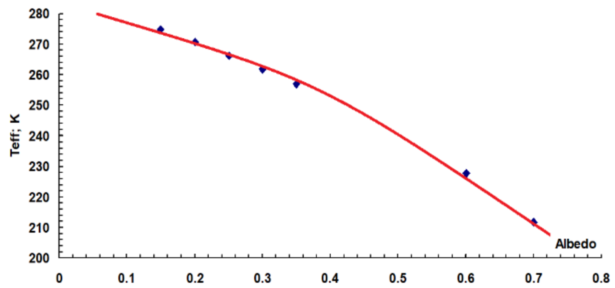

Figure 26: The values of effective temperatures as the function of Albedo (based on the data from the Table 9)

The known values of effective temperatures T_{eff} from the Table 7 allow again the recalculation of the values of the energies of outgoing longwave radiation E_{th} and to compare with the real ones. The representation of the dependence $E_{th}=f(E)$ is represented on Figure 27. The correlational calculations show the strong link between data and the calculated values of E_{th} are of the same order as *E*.

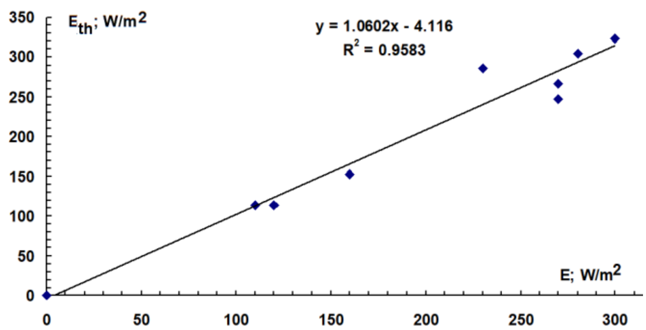

Figure 27: The reciprocal interconnection of the data *E* and E_{th} (based on the data from the Table 7)

According to Wien's displacement law the wavelengths are calculated (Table 9). The dependence of λ=f(A) is represented on Figure 28.

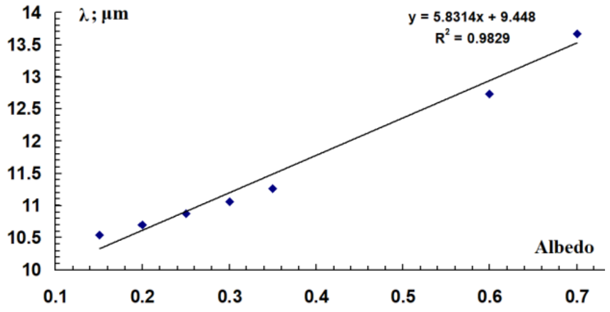

Figure 28: The dependence of wavelengths of outgoing radiation as the function of Albedo (based on the data from the Table 7)

The cooling of Earth's upper atmosphere by the increasing of the pollutants green house gases is also described recently in [43]: „The results confirmed that rising carbon dioxide levels were the main driving force cooling the upper atmosphere". In such way, the quantitative explanation by the expressions of albedo is performed and the validation of results is checked by real recent examples presented in various studies.

Conclusions

The physical-mathematical description of the ecological, climatic and microclimate state of the geographical regions remains always one of the main task of the study. The answer to the question of how the albedo value influences the determination of the respective equilibrium temperature of the components generally is solved.

When talking about the albedo value, it is important to note that this term is used for the full atmosphere-Earth system. it is observed that with the increase of the atmospheric temperature due to of the green house effect the albedo value of this system increases.

The empirical expression of albedo which is obtained in this study depends on the temperature of the atmosphere, the Earth's heat content, the amounts of the masses of the decline in oxygen and the accumulation of carbon dioxide.The value of thermal emission of Earth is checked by real values of the registration. The respective wavelengths of ongoing longwave radiation are increasing with the increasing of albedo values. This is shown by the example of upper atmosphere when it is cooling by the increasing of greenhouse pollutants.

The practical importance of the suggested method consists in the application of the quantitative calculation of the effective temperature for various geographical regions by the known value of the emitted energy which is measured by technical means. In its turn, the measured value of the energy by technical means of the thermal radiation of the earth (W/m^2) allows to calculate the albedo values for the respective geographical place and then can be compared with the real albedo that can be measured by technical means. It is clear that the values of effective temperature, thermal radiation energy and albedo vary from one geographical location to another. This quantitative relation of the effective temperature can be successfully applied to the process of measuring the thermal energy (W/m^2) emitted by the respective geographical place.

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