INFLUENCE OF WATER AND WATER-ACID AEROSOLS ON TROPOSPHERE ENERGY BALANCE DURING SUN-PROTON EVENTS (SPE) IN THE HIGH-ALTITUDE AREAS

Aleksey L. Kovorotniy, Yuriy V. Goncharenko, and Vladimir N Gorobets

Usikov Institute of Radiophysics and Electronics, National Academy of Sciences of Ukraine,
12, Proskura str., Kharkov, 61085, Ukraine,
email: kovorotniy@ire.kharkov.ua

Abstract

*This paper considers influence of water and water-acid aerosol layers formed during s*olar *proton events (SPE) on the atmosphere energy balance in the height latitude region. It is demonstrated that flux of solar energy reaching the Earth can decrease by 8-15% due to reflection from aerosol layers of different types and formed as a result of ionizing influence of high-energy solar protons.*

1. INTRODUCTION

It is known that increased activity of the Sun can have great influence on weather and climate of the Earth [1, 2]. This effect is based on numerous experimental facts, which establish correlations and other relationships between indicators of solar activity, such as Wolf numbers, sunspot number and time of their crossing through the central meridian, recurrent events in the atmosphere, connected with Sun rotation and others, connected with meteorological parameters of the atmosphere [1].

There was made a suggestion in works [2, 3] that one of the mechanisms changing atmospheric parameters during solar proton events (SPE) is formation process of water and water-acid aerosol on charged particles. However, considered models don't include estimates of influence of such aerosols taking into account their concentration, chemical composition and size distribution.

The purpose of the present work is to get numerical estimate of influence of aerosol formation on atmosphere energy balance.

2. THE OPTICAL MECHANISM INFLUENCES OF SPE ON THE CHARACTERISTIC OF MEAN AND BOTTOM ATMOSPHERE

Work [4] demonstrates data of temperature change at different altitudes during SPE. These data were received at Sodankyla station (Finland). This work shows that

during an SPE the temperature of troposphere is increasing at altitudes h<8 km and decreasing at altitudes h>8 km. This may result from its screening by aerosol formations. Modeling presented in the work [4] has shown that received experimental data on temperature profile alteration can be explained by inclusion of layer with reflection coefficient R=0.1 at altitudes 8-10 km (Figure 1). As increased solar activity can lead to increase of ion concentration at altitudes 10-15 km from 100 cm^{-3} to 1000 cm^{-3} [1] it can be supposed that aerosol concentration at these altitudes is approximately by one order of magnitude greater.

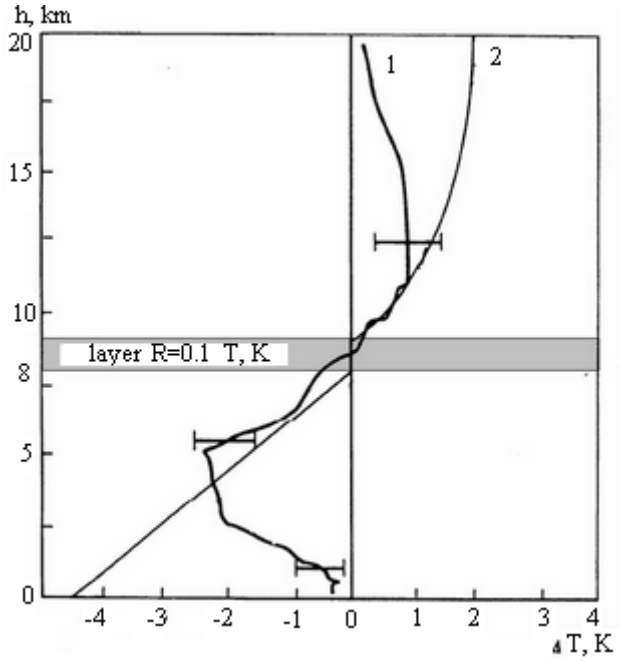

Figure 1. Temperature profile change of lower and middle troposphere: 1 – experiment; 2 – model with reflecting layer at altitudes 8-9 km

The integral attenuation coefficient of solar radiation in the troposphere aerosols depends on many factors, including nature of aerosol, particle size and its concentration.

3. ATMOSPHERIC AEROSOL TYPES

From all the diversity distributions of droplets and particles in atmosphere, paper [5] considers six models with regard to the particle size distribution (haze M, L, H; rain M, L and deg H).

First three models are distinguished to describe aerosol particles. Thus, M haze is typical for characteristic of marine aerosols. Haze L is typical for aerosols of continental origin. Haze H describes distribution of high altitude and stratospheric water aerosol.

Other constituents of the atmosphere, such as compounds of nitric or sulfuric acid, may also be presented as a component of atmospheric aerosol. According to Dickinson [6], nuclei of condensation are represented by atmospheric clusters formed during the ionization, and the main atmospheric aerosol – by a mixture of sulfuric acid and water.

Figure 2 shows distribution functions of aerosol particles by size for marine, mineral, water and water-acid aerosols. The distributions 3 and 4 are of outmost interest when analyzing the effect of SPE on the energetic atmosphere balance.

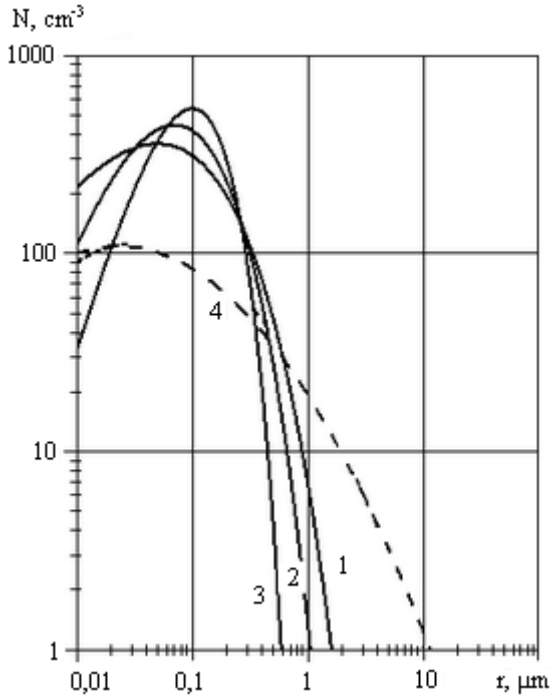

Figure 2. Function of haze particle size distribution: 1- haze M, 2 – haze L; 3 – haze H; 4 – water-acid aerosol;
—— Dermidgian distribution; – – – logarithmically normal distribution law

Let us assume that increased solar activity can increase concentration of ions and other particles which in their turn represent condensation nuclei for water and water-acid aerosols. Let's consider scattering and absorption of electromagnetic energy on aerosol particles of these types.

In the studied range of wavelengths dielectric capacitance of pure water and acid-water aerosols has a small imaginary part so we can consider only mechanism of scattering of plane electromagnetic waves from drops, neglecting absorption.

3. QUANTITATIVE ESTIMATE OF AEROSOL FORMATIONS INFLUENCE ON ENERGY BALANCE OF ATMOSPHERE

It is known that 98% of solar energy stands in wavelength range from 0.2 to 3 μm [7] and this energy is approximately constant in time. So, we should principally consider atmosphere aerosols influence on energy transfer in this wavelength range.

The integral coefficient of attenuation of solar radiation in wavelength range from 0.2 to 3 μm, introduced by layer with aerosol distribution from 0.005 to 2.66 μm for water-acid aerosol layers, and from 0.07 to 1.4 μm for water aerosol layers, can be determined by the formula (1):

$$\alpha_{att}(r) = \frac{1}{W_0} \int_{r_1}^{r_2} \int_{\lambda_1}^{\lambda_2} \Gamma_{scat}(\frac{2\pi r}{\lambda}, m) \cdot W(\lambda) d\lambda dr \qquad (1)$$

where W_0 is the solar constant equal to 1373±20 W/m^2 [7]; Γ_{scat} is the scattering coefficient; r is the drop radius (μm); λ is the wavelength range (μm); m is the complex dielectric constant [8]; $W(\lambda)$ is the solar energy spectrum.

Scattering coefficient can be determined by Buger law [9]:

$$\Gamma_{scat} = 1.346439 \cdot 10^{-2} \int_{r_1}^{r_2} r^2 N f(r) K_{scat}(m, x, \theta_1, \theta_2) dr \qquad (2)$$

where N is the number of drops in 1 cm^{-3}; $f(r)$ is the density of drops distribution by size; K_{scat} $(m, x, \theta_1, \theta_2)$ is the coefficient of efficiency distribution at the angle range $\theta_1..\theta_2$; $x = kr$ is the wavelength constant; k is the wave number.

Dermidgian has proposed three basic models of atmospheric aerosols to describe water, continental and stratospheric water aerosols:

$$f(r) = A_1 r^{\alpha_1} \exp[-b_1 r^{\gamma_1}] \qquad (3)$$

where A_1, b_1, α_1 and γ_1 are the constants, positive real numbers associated with each other and characterizing density of distribution of particle size .

In the work [3], to describe water-acid aerosol particle distribution it was suggested to use a lognormal law of distribution:

$$f(r) = \frac{1}{\sqrt{2\pi} \cdot \sigma \cdot r} \cdot \exp^{-\frac{(\ln r - d_\Gamma)^2}{2 \cdot \sigma^2}} \qquad (4)$$

where σ is the standard geometric deviation equal to 1.89; d_Γ is the average particle diameter equal to 0.139 μm.

Coefficient of energy scattering efficiency will be calculated in the angle range $\theta_1..\theta_2$ with taking into account the ratio of relation of energy dispersed into rear hemisphere to total energy affecting the particle:

$$K_{scat}(m, x, \theta_1, \theta_2) = k_i(\theta_1, \theta_2) \cdot k_{scat}(m, x) \tag{5}$$

where, k_{scat} is the scattering efficiency; $k_i(\theta_1, \theta_2)$ is the coefficient taking into account the ratio of fluxes of energy scattered by particle in the angle range $\theta_1..\theta_2$.

Let us take into account that sizes of aerosol particles are comparable with wavelengths therefore Mie equations can be used for calculation of attenuation parameters introduced by aerosol particles [5].

Scattering efficiency of one drop is calculated by formula (6):

$$k_{scat}(m, x) = \frac{2}{x^2} \sum_{n=1}^{\infty} (2n+1)(|a_n|^2 + |b_n|^2) \tag{6}$$

where a_n and b_n are Mie scattering coefficients [5]; n are positive integer numbers.

Intensity of radiation dispersed by aerosol particles of spherical shape will be equal to [10] in some points at distance r' from the particle:

$$I(\theta) = \frac{I_0(i_1(x, m, \theta) + i_2(x, m, \theta))}{2k^2 r'^2} \tag{7}$$

where I_0 is the intensity of falling non-polarized radiation; i_1, i_2 are the parameters of intensity.

Ratio of fluxes of energy dispersed by particle in the angle range $\theta_1..\theta_2$ is calculated as:

$$k_i(\theta_1, \theta_2) = \frac{\int_{\theta_1}^{\theta_2} I d\theta}{\int_0^{2\pi} I d\theta} = \frac{\int_{\theta_1}^{\theta_2} [i_1(x, m, \theta) + i_2(x, m, \theta)] d\theta}{\int_0^{2\pi} [i_1(x, m, \theta) + i_2(x, m, \theta)] d\theta} \tag{8}$$

Parameters of intensity i_1, i_2 are related with complex scattering amplitudes $S_1(\theta)$ and $S_2(\theta)$ and describe in [10].

The results of numerical calculation of the attenuation coefficient (1) for solar radiation by aerosol layer with thickness 0.5 km [11] were calculated for two cases: for quiet atmosphere when the concentration of aerosols does not exceed the background values ($N \leq 10^2$ cm^{-3}) and during SPE when concentration can be increased up to 10^3 cm^{-3} (Figure 3).

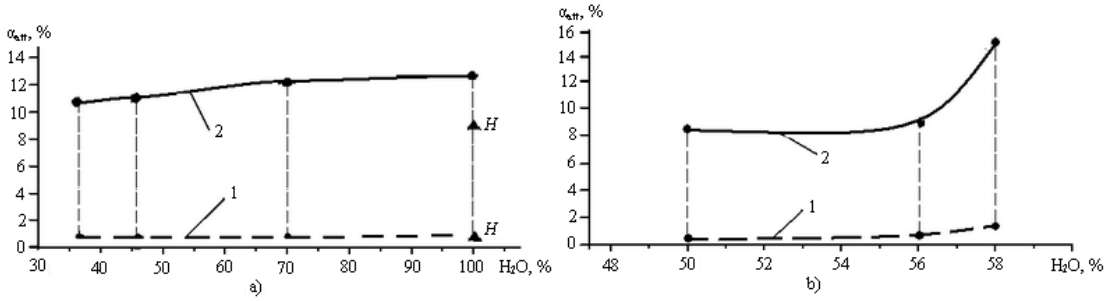

Figure 3. Coefficient of electromagnetic energy attenuation by aerosol layer during SPE: a) two-component mixture (H_2O, HNO_3); b) three-component mixture (H_2SO_4, HNO_3, H_2O)

It is shown that attenuation of solar radiation on the aerosol particles of water and water-acid origin does not exceed 1.5% during the quiet atmosphere (curve 1 in figures 3a and b) and does not have significant effect on energy balance of troposphere. During powerful SPEs (curve 2) concentration of aerosol particles can significantly increase and exceed baseline values in dozens of times. Amount of energy reflected into upper hemisphere can reach up to 8-15%. As a result, the air air will be heated up in the upper troposphere and lower stratosphere, but consequently the cooling air is under the aerosol layer.

It should be noted that water aerosol with "*H*" distribution (point *H* in figure 3a) has a minimum influence on electromagnetic energy transfer. That is when a layer consists only aerosol water particles with the distribution coinciding with the distribution of high altitude haze, the amount of energy reflected into upper atmosphere does not exceed 0.8% (quiet conditions) and may reach 8.5% during SPE. The influence of water aerosol with logarithmically normal distribution (point with 100% humidity) is also not important: 1% before and 12% after SPE.

Mixture consists of three components may have the biggest impact on energy balance of troposphere. The weakening of solar energy during the SPE can reach 15% for three-component mixture $H_2SO_4/HNO_3/H_2O$ = 28/14/58%.

The number of solar protons penetrated in Earth atmosphere directly depends on the sun elevation at horizon. The figure 4 shows dependence of attenuation coefficient of solar energy on aerosol layers consisted of different type aerosols from season.

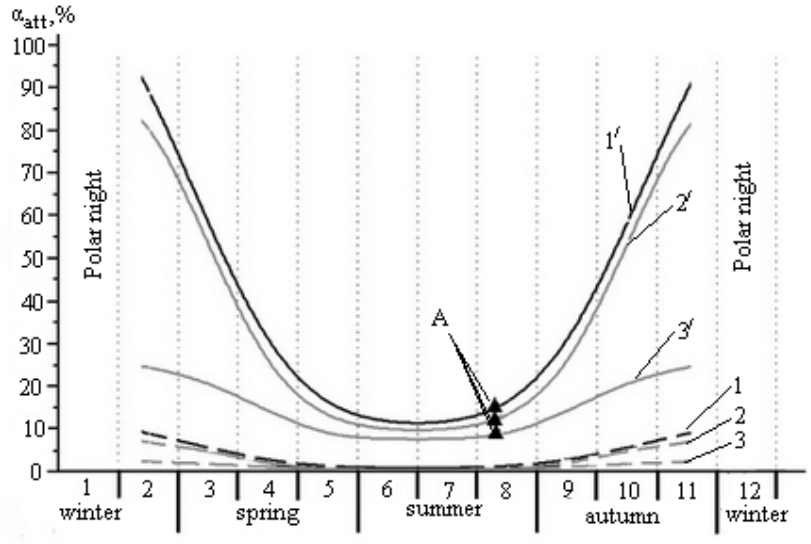

Figure 4. The dependence of attenuation coefficient of solar energy to aerosol layers from season: A - attenuation coefficient during SPE (August 4, 1972); − − − - quite atmosphere (1 – H_2SO_4; 2 – HNO_3; 3 – H_2O); ——— - during SPE (1′ – H_2SO_4; 2′ – HNO_3; 3′ – H_2O)

The data presented on the figure 4 are the result of mathematical modeling. The calculation was performed for Sodankyla station (coordinates: 67°22' N, 26°39' E) and taken into account two cases: for quiet atmosphere (curve 1, 2 and 3 on the figure3) and during SPE (curve 1′, 2′ and 3′). Solar proton events which occurred on August 4, 1972 were taken as a basis for this model scheme [12].

4. CONCLUSIONS

Thus, during the powerful SPE increased level of ionization of lower stratosphere and top troposphere in high altitude regions may increase the condensation nuclei number for atmosphere aerosol in 5-10 times and lead to formation of aerosol layers with high density.

It is shown that such layers can lead to attenuation of solar radiation flux near Earth's surface to value reaching 15%. This does not conflict with data of modeling of Earth's atmosphere energetic balance the results of which show that amount of energy reflected into upper atmosphere layers can reach 10-15% of total solar radiation.

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