AN ACCOUNT OF ATMOSPHERIC CHANNEL IN PROCESSING OF THE SCANNOGRAMS OF A GLOW IN THE UPPER ATMOSPHERE

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The effect of measurement geometry and atmospheric extinction on scannograms of background radiation of the upper atmosphere and aurora is investigated. The atmospheric extinction due to Rayleigh and Mie scattering has been calculated for several statistically well–founded models of the aerosol atmosphere. It is found that the McClatchey model ($S_{\rm m}$ = 50 km) modified in the lower 4–kilometer layer to take into account the anthropogenic effects is the best one for measurements carried out in Noril'sk. It is established that elimination of the effects connected with the measurement geometry and atmospheric extinction from the aurora scannograms results in significant changes of their structure.

At present a study of interconnection in a system "lithosphere–ionosphere–magnitosphere" of the Earth becomes urgent. A great number of experimental data, testifying considerable influence of Earth's crust structures anysotropic in their physical properties, on spatial distribution of aurora borealis have been stored (terrigeneous effect).[1–3] This subject matter is urgent due to the fact that elucidation of the physical mechanism of terrigeneous effect allows us to develop optical remote methods of hypogene geological mapping at high latitudes and to refine the model of the polar atmosphere.

At present the main problem is to relate experimentally the structure of the earth crust and spatial distribution of aurora borealis. It is necessary since the experimental data currently available are insufficient and ambiguous, and a unified technique for processing of ground–based photometric measurements of aurora borealis is lacking. One of the most important factors affecting ambiguity of mapping of the contours of aurora recurrence is the lack of unified model for the optical atmospheric parameters, which can be used for processing of experimental data, and of the correct technique which allows for a night self–glow of the atmosphere.

The present paper deals with the problems connected with allowance for the influence of viewing geometry and atmospheric extinction on scannograms of a night self–glow of the atmosphere and aurora borealis.

Background radiation is modeled by a spherical radiating layer located at the altitude h from the Earth's surface and having the thickness Δh (Fig. 1). The model is elaborated for a scanning photometer with the field–of–view angle of 2ω degrees. The photometer records the radiation coming from the volume $V(\theta)$ of the radiating layer. This volume is calculated from the equation

$$V(\theta) = 1/3\ \pi \tan^2(2\omega)\left[r_2^{\,3}(\theta) - r_1^{\,3}(\theta)\right] . \qquad (1)$$

Here

$$r_1 = -R\cos(\theta) + [R\cos(\theta) + 2R\,h + h^2]^{1/2} ,$$

$$r_2 = -R\cos(\theta) + [R\cos(\theta) + 2R\,(h + \Delta h) + (h + \Delta h)^2]^{1/2} .$$

Radiant power arriving at the photometer can be represented in the form

$$P(\theta) \sim \frac{h\,\nu}{\tau_l}\,N(\theta)\ V(\theta)\ T(\theta) , \qquad (2)$$

where $N(\theta)$ is the number of excited radiating centers per unit volume, τ_l is the lifetime of the center in excited state, $T(\theta)$ is the slant transparency of the atmosphere, h is Planck's constant, and ν is the radiation frequency.

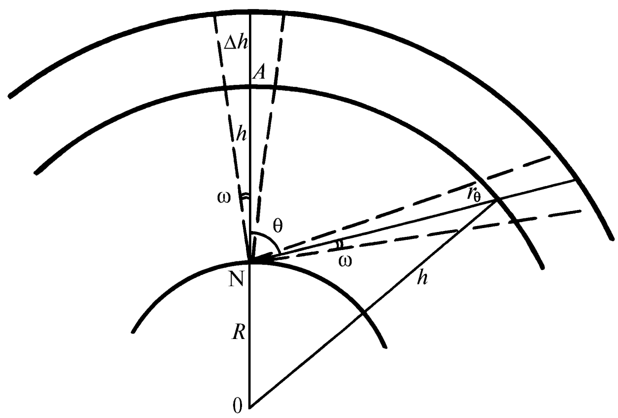

FIG. 1. Radiating layer model. Here h is the layer altitude, Δh is the layer thickness, θ is the zenith angle, ω is the field–of–view angle of a photometer, R is the Earth's radius, and N is the point at which the photometer is placed.

It is assumed that quenching of radiation by foreign particles is independent of θ, i.e., within the field of view of the photometer the atmosphere is horizontally homogeneous. The parameter $T(\theta)$ is calculated by the formula

$$T(\lambda,\,\theta) = \exp\left(-\int\limits_0^{r_\theta} \alpha\,(\lambda,\,r)\,{\rm d}r\right) . \qquad (3)$$

Since in the atmospheric models only the vertical distribution of the extinction coefficient is assigned, the above integrand is calculated as a function of altitude

$$z(r_\theta,\,\theta) = \sqrt{R^{\,2} + r^{\,2} + 2\,R\,r\cos(\theta)} - R .$$

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The volume extinction coefficient $\alpha(z, \lambda)$ can be represented in the form

$$\alpha(z, \lambda) = \beta_m(z, \lambda) + \alpha_a(z, \lambda) + K_m(z, \lambda) , \quad (4)$$

where $\beta_m(z, \lambda)$ is the Rayleigh scattering coefficient, $\alpha_a(z, \lambda)$ is the aerosol extinction coefficient, and $K_m(z, \lambda)$ is the molecular absorption coefficient.

The values of the Rayleigh scattering coefficients near the Earth's surface at the pressure P_0 = 1013 mb and the temperature T_0 = 247 K were calculated according to the formula[4]

$$\beta_m(z_0, \lambda) = \frac{8\pi^3(n^2-1)^2}{3N\lambda^4}\frac{6+3\delta}{6-3\delta} , \quad (5)$$

where $\delta = 0.035$ is the depolarization ratio, N is the number of molecules per unit volume, and n is the refractive index defined by the Edlen equation[5]

$$(n-1)\cdot 10^6 = 64.328 + 29498.1/(146-\lambda^{-2}) + 255.4/(41-\lambda^{-2}). \quad (6)$$

Here λ is the wavelength in μm.

The vertical distribution $\beta_m(z, \lambda)$ was calculated according to the formula

$$\beta_m(z, \lambda) = \beta_m(z_0, \lambda)\frac{P(z)\,T_0}{P_0\,T(z)} , \quad (7)$$

where $P(z)$ is the vertical pressure profile and $T(z)$ is the vertical temperature profile, needed for the calculations. These profiles for the atmospheric model at polar latitudes in winter are given in Table I (see Ref. 6).

In calculation of the transparency, the possibility of molecular absorption for the given wavelength should be analyzed, i.e., the magnitude of K_m should be evaluated. Thus, at $\lambda = 0.5577$ μm such optically active gaseous constituents as H_2O, NO_2, and O_3 may absorb. The fine spectrum of water vapor in the frequency region 16500–25250 cm^{-1} was measured in Ref. 7. As seen from these measurements, the radiation at $\omega = 0.5577$ μm and 17930.78 cm^{-1} is in the transparency windows: the nearest rovibrational transitions are centered at $\omega = 17764.75$ and 18066.89 cm^{-1}. The NO_2 molecule in the considered spectral region has continuous spectrum with pronounced vibrational structure. According to the data reported in Ref. 8, the cross section of absorption $\sigma_{NO_2}(0.5577) \simeq 6.3\cdot 10^{-20}$ cm^2; however, low volume NO_2 concentrations (~10 ppb at a maximum reached at an altitude of 30 km) make the influence of this absorption negligible. There are several weak diffuse Chappeu bands in the absorption spectrum of ozone in the spectral region from 0.5500 to 0.6020 μm. According to the data of Ref. 9, $\sigma_{O_3}(0.5577)$ is approximately equal to $3.2\cdot 10^{-21}$ cm^2. To estimate the contribution of ozone to total absorption, we calculated the transparency T at $\lambda = 0.5577$ μm and $\theta = 0°$ from formula (3) without and with regard to the absorption by ozone. In the first case the molecular absorption coefficient $K_m(z, \lambda)$ was assumed to be equal to zero in Eq. (4). In the second case $K_m = 2.32\cdot 10^{-5} P_3\Delta H$, where P_3 is the ozone pressure, in nanobars and ΔH is the step of integration over the altitude, in km. The values of P_3 were chosen from the model distribution of ozone concentration at polar latitudes in winter.[6] The values $T = 0.75$ and 0.67 were obtained without and with regard to the ozone absorption. These values are indicative of ozone important contribution to the results of quantitative calculations.

TABLE I. Vertical profiles of atmospheric pressure and temperature for polar latitudes in winter.

H, km	P, mb	T, K
0	1.013+3	247
1	8.878+2	251
2	7.775+2	249
3	6.798+2	245
4	5.932+2	240
5	5.158+2	234
6	4.467+2	225
7	3.853+2	223
8	3.308+2	219
9	2.829+2	216
10	2.418+2	216
11	2.067+2	216
12	1.766+2	216
13	1.510+2	216
14	1.291+2	216
15	1.103+2	216
16	9.431+2	215
17	8.058+2	215
18	6.882+2	214
19	5.875+2	213
20	5.014+1	212
21	4.277+1	212
22	3.647+1	211
23	3.109+1	211
24	2.649+1	211
25	2.256+1	212
26	1.925+1	212
27	1.642+1	212
28	1.401+1	212
29	1.195+1	212
30	1.020+1	212
35	4.701	222
40	2.243	235
45	1.113	247
50	5.719–1	259
55	2.944–1	256
60	1.515–1	253
65	7.801–2	249
70	4.016–2	246
80	7.850–3	234
90	1.535–3	222
100	3.000–4	210

The choice of aerosol model in the calculation of atmospheric optical model is difficult due to wide variability of concentration, microstructure, and chemical composition of aerosols at different altitudes. According to the McClatchey model,[18] reflecting such a variability, the atmosphere is devided into four layers: boundary layer (0–2 km), upper troposphere (2–10 km), low stratosphere (10–30 km), and upper atmosphere (30–100 km). The boundary layer is described by the widest variety of situations. Continental, marine, tropospheric, and urban models of aerosol are used here, each for the meteorological visibility range S_m = 50, 23, 10, 5, and 2 km at the relative humidity U = 0, 70, 80, and 99%. For the upper troposphere, the model duplicates the continental model of the boundary layer with high values of S_m (23 and 50 km). For the low stratosphere four models of aerosols are accepted, namely: background as well as moderately, highly, and extremely volcanic aerosol models. For the upper atmosphere the hypothesis of meteorite dust has been accepted.

In our calculations we started from the model of background tropospheric and stratospheric aerosols with $S_m = 50$ km and normal upper atmosphere. Along with this model[10] the background and mean–cyclic models of aerosol proposed by G.M. Krekov et al.[11] was used. The vertical profiles of aerosol scattering coefficients for these models are shown in Fig. 2 for a wavelength of 0.55 μm.

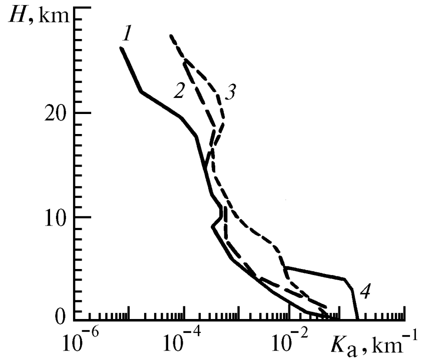

FIG. 2. Vertical profiles of the aerosol scattering coefficients for different models of the aerosol atmosphere: 1) background model, 2) mean–cyclic model, 3) McClatchey model ($S_m = 50$ km), and 4) refined McClatchey model ($S_m = 50$ km).

Atmospheric radiation intensity recorded by a photometer can be represented in the form

$$I(\theta) = K\, V(\theta)\, T(\theta)\ , \tag{8}$$

where K is the calibration constant. For its determination we write down relation (8) for the experimental value of the radiation intensity measured in the zenith direction ($\theta = 0$)

$$I_{0\mathrm{exp}} = K\, V_0\, T_0\ , \tag{9}$$

where $V_0 = V(0)$ and $T_0 = T(0)$ are the radiating volume and the transparency in the zenith direction. It follows from Eq. (9) for the calibration constant

$$K = I_{0\mathrm{exp}} / V_0\, T_0\ .$$

After substitution of the constant K into Eq. (8) we obtain the equation for the calculation of model background scannograms when the results of experimental measurements are available

$$I(\theta) = I_{0\mathrm{exp}} \frac{V(\theta)\, T(\theta)}{V_0\, T_0}\ . \tag{10}$$

In the above reasoning the constant K was assumed to be independent of the angle θ, i.e., the radiating layer was assumed to be horizontally homogeneous. The truth of this hypothesis was varifield by us with the use of the MSIS–86 tropospheric model of the atmosphere.[12] With the help of such a model the variations in the concentration of atomic oxygen (radiating agent) and molecular oxygen (quenching agent) for the average altitude of aurora borealis $H = 110$ km were evaluated for the point of location of the photometer placed 100 km north of Noril'sk. Scanning was performed along the meridian, so that the segments of the radiating layer in the 68–72°N latitude belt fell within the field of view of the photometer. Under such conditions, according to the MSIS–86 model, the variations in the concentration of atomic oxygen did not exceed 3%, and that for molecular oxygen – 0.3%. An example of processing of experimental scannogram by the given technique in the absence of aurora borealis is shown in Fig. 3 for a wavelength of 0.5577 μm. Experimental scannogram was obtained on February 13, 1991 at 13:20 h, UT for a visibility range corresponding to a cloud amount of 5, according to Ref. 13. As seen from Fig. 2, models 1–2 deviate from the real distribution of aerosol components as does the McClatchey model ($S_m = 23$ km). The McClatchey model for $S_m = 50$ km is practically identical to the mean–cyclic model.[11]

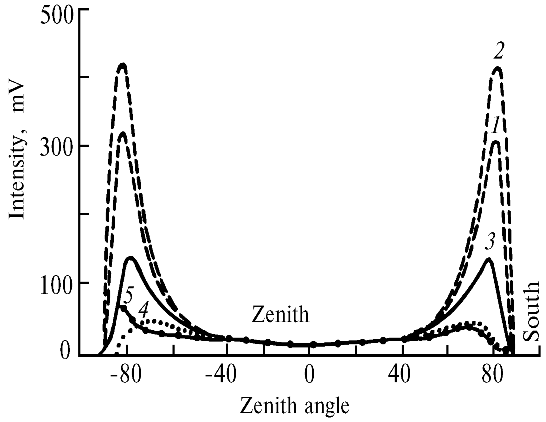

FIG. 3. Experimental and model scannograms for the 5577Å *line and different models of the vertical profiles of aerosol components of the atmosphere: 1) mean–cyclic model according to Ref. 11, 2) background model according to Ref. 11, 3) McClatchey model* ($S_m = 23$ km) (*see Ref. 10*), *4) McClatchey model* ($S_m = 50$ km) *refined at altitudes up to* 5 km, *and 5) experimental scannogram recorded on February 18, 1991 at* 13:20 h, UT.

Disagreement between the experiment and model may be caused at least by two reasons. First, by imperfection of the models themselves due to the lack of the well–founded data on concentration and optical properties of aerosols near the polar latitude. The second reason may be connected with the possible local anthropogenic effect within the observation zone. In our case a major source of such an influence may be Noril'sk mining–metallurgical plant. According to the official data, the amount of emission of polluting substances from this plant may reach 2368 million tons per year.[14]

Considering the anthropogenic factor as the main reason of discrepancies, we refined the McClatchey model ($S_m = 50$ km) at altitudes up to 4 km (curve *4* in Fig. 3) against the criterion of best fitting the experimental data. The standard deviation of the intensity of model scannogram from the experimental one was 1.068. The vertical distribution of the aerosol scattering coefficients is shown by curve *4* in Fig. 2. The above–described model was used by us to correct two experimental scannograms shown in Fig. 4. These scannograms were used in calculation of the normalized concentrations $N(\theta)$ of excited particles via the experimental intensities $I_{\mathrm{exp}}(\theta)$ from the formula

$$N(\theta) = \frac{I_{\mathrm{exp}}(\theta)}{V(\theta)\, T(\theta)}\ .$$

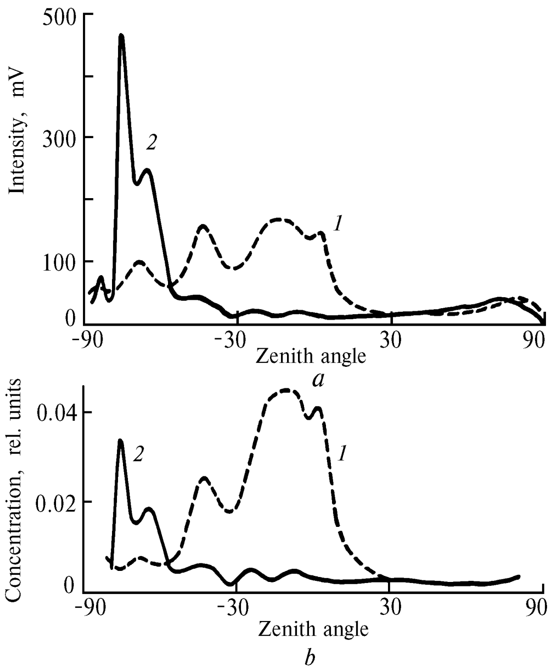

FIG. 4. Scannograms of aurora borealis at a wavelength of 5577Å *(a): 1) February 13, 1992 at* 20:00 h, UT, a_1SA1cz *and fRB1cN2 aurora borealis, 2) February 18, 1992 at* 16:55 h, UT, *RB2cN2 aurora borealis. Angular distribution of the concentration of radiating centers (b).*

The results of calculations are shown in Fig. 4*b*. It is seen from this figure that the segments of scannograms calculated at large zenith angles both north and south the zero angle are primarily subject to changes. At small and moderate zenith angles the spatial structure of scannograms retains its shape, but relative intensities of individual components can significantly change. An account of such changes is essential in statistical processing of the experimental scannograms.

CONCLUSION

Thus specifying the radiation wavelength and altitude and thickness of the radiating layer, one can model the scannogram in the absence of aurora borealis and correct it for the geometry of observations and the atmospheric extinction. Disadvantages of the model are neglect of the following factors:

– multiple scattering on near–surface paths,

– diffuse reflection from the Earth's surface,

– disagreement between real and model distributions of aerosol components.

Although the diffuse scattering from the underlying surface can be easy taken into account by introducing the albedo of the Earth's surface, certain difficulties arise with the remaining factors, since an account of a great number of parameters is required.

Starting from the above–mentioned, we conclude that this model is applicable at the zenith angles up to 70° at which the error due to multiple scattering for the considered wavelengths does not exceed 20% (see Ref. 15). The experimental data indicate that in Noril'sk region the McClatchey model (S_m = 50 km) describes well the model of polar atmosphere with allowance for changes in the boundary layer, i.e., during photometric measurements the distribution of the aerosol scattering coefficient at altitudes up to 5 km should be monitored and the visibility range should be measured.

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