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Solutions of some retrieval problems on the basis of ground actinometry

E. M. FEIGELSON

Institute of Atmospheric Physics, Russian Academy of Sciences. Russia

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RESUMEN

Se demuestra la posibilidad de determinar algunos parámetros de importancia óptica y otros parámetros con base en actinometría estándar en la superficie.

Esos parámetros son: espesor óptico de una atmósfera libre de nubes, espesor óptico de cubierta total de nubes, cantidades de nubes separadas y el flujo total de radiación térmica saliente.

ABSTRACT

The possibility of determination of some important optical and other atmospheric parameters on the basis of standard ground actinometry is shown.

These parameters are: aerosol optical thickness of the clear atmosphere, optical thickness of overcast clouds, broken cloud amounts and the total flux of the outgoing thermal radiation.

Introduction

In atmospheric radiation and cloud problems concerning climate studies most attention now is on the retrieval of satellite measurements. But we have a world-wide continuously operating ground actinometry network. Many important retrieval results may be obtained from these measurements.

Some examples are given below, based on the author's and her colleagues work.

1. Determination of the spectral aerosol optical thickness on the basis of measurements of total direct solar radiation

This method was proposed by Krasnokutskaya and Feigelson (1980) and made more precise and detailed by Tarasova and Yarcho (1991).

Total direct solar radiation, I, measured on the ground may be expressed in the form:

$$I = \int_{0.2}^{4.0} I_{0,\lambda} \exp[-\sec\xi(\tau_{0z,\lambda} + \tau_{R,\lambda} + \tau_{w,\lambda} + \tau_{a,\lambda})]d\lambda \qquad (1)$$

Here $\tau_{0z,\lambda}$, $\tau_{R,\lambda}$, $\tau_{w,\lambda}$, $\tau_{a,\lambda}$ are the spectral optical thickness due to ozone absorption (0_3), Rayleigh scattering (R), water vapour absorption (W), aerosol extinction (a); ξ is the zenith angel of the sun; λ in μm.

The spectral distribution of $\tau_{a,\lambda}$ is described in the following way:

$$\tau_{a,\lambda} = \tau_{a,\lambda_o}(\lambda_o/\lambda)^n \text{ at } \lambda_o = 0.55\mu m \qquad (2)$$

and

$$\tau_{R,\lambda} = 0.098(\lambda_o/\lambda)^4 \qquad (3)$$

Calculations of I in accordance with equation (1) were carried out in the range of parameters: $0 \leq \xi \leq 75^o$; $0 \leq W$ g/cm$^2 \leq 5$; $0.01 \leq \tau_{a,\lambda_o} \leq 1.0$; $0 \leq n \leq 2$; here W is the total water vapour content.

$\tau_{0z,\lambda}$ was taken as constant, corresponding to an ozone content equal to 0.3 cm. In this case $\tau_{0z,\lambda}$ is distributed in the following way:

Table 1. Spectral optical thickness of ozone got in accordance with extinction coefficients represented in "Global Ozone Research" (1985)

$\Delta\lambda, \mu$m =	0.29-0.30;	0.30-0.32;	0.32-0.34;	0.34-0.36;
$\tau_{0z,\lambda}$ =	6.21	0.65	0.034	0.0004
$\Delta\lambda, \mu$m =	0.44-0.48;	0.48-0.52;	0.52-0.56;	0.56-0.60;
$\tau_{0z,\lambda}$ =	0.003;	0.0099	0.037;	0.040;
$\Delta\lambda, \mu$m =	0.60-0.64;	0.64-0.68;	$\lambda > 0.68$	
$\tau_{0z,\lambda}$ =	0.032;	0.018;	0	

A shortened version of the got tables of I (see Tarasova and Yarcho, 1991) is given in Table 2:

Table 2. The calculated total direct solar radiation I kw/m for $\tau_{a,\lambda_o} = 0.1$ and 0.5; $\xi = 60^o$, different n and W.

W g/cm^2	n = 0		0.5		1.0		2.0	
τ_{a,λ_o}	0.1	0.5	0.1	0.5	0.1	0.5	0.1	0.5
0	0.926	0.416	0.961	0.510	0.984	0.587	1.008	0.691
0.5	0.815	0.366	0.841	0.435	0.859	0.493	0.877	0.575
1.0	0.792	0.356	0.816	0.420	0.833	0.475	0.850	0.551
2.0	0.766	0.344	0.789	0.404	0.804	0.455	0.820	0.526
3.0	0.750	0.337	0.772	0.394	0.786	0.442	0.801	0.511
5.0	0.729	0.328	0.750	0.381	0.763	0.426	0.777	0.491

The imput of the measured I in a table like 2 but more detailed permits to find the value or τ_{a,λ_o}, if w, ξ and n are given. For $n = 1$ (which may be considered as a mean value) a formula is given (see Tarasova and Yarcho, 1991):

$$\tau_{a,\lambda_o} = \frac{\ln I - [(0.1886W^{-0.1830} + (0.8799W^{-0.0094} - 1)/\cos\xi]}{0.8129W^{-0.021} - 1 + (0.4347W^{-0.0321} - 1)\cos\xi} \tag{4}$$

Table 2 shows that dependence of I on W is more important than of n. There are two ways to get W: 1) To measure it directly, for instance by a ballon. 2) To use the empirical correlations between W and e_o -the water vapour pressure near the surface. Such correlations are presented in the book Abakunova *et al.* (1983) for different places of the former Soviet Union.

For Moscow we get:

$$W = 1.6e_o + a \tag{5}$$

with $a = 2.3$ for winter, 1.3 for spring, 1.6 – for summer, 2.1 – for autumn; e_o in millibars; W in millimeters.

2. Determination of the cloud optical thickness on the basis of the global radiation measurements

The measured surface fluxes of the global radiation in the visual range of waves length– $Q_{vis}(0.38 \leq \lambda \leq 0.70\mu\text{m})$ depend mostly on the cloud optical thickness τ and the zenith angle of the sum ξ. If the cloud type is known (assumed to be homogeneous and overcast), the same as the corresponding effective radius of the particle $-r_{ef}$, the Q_{vis} dependence on τ and ξ can be calculated.

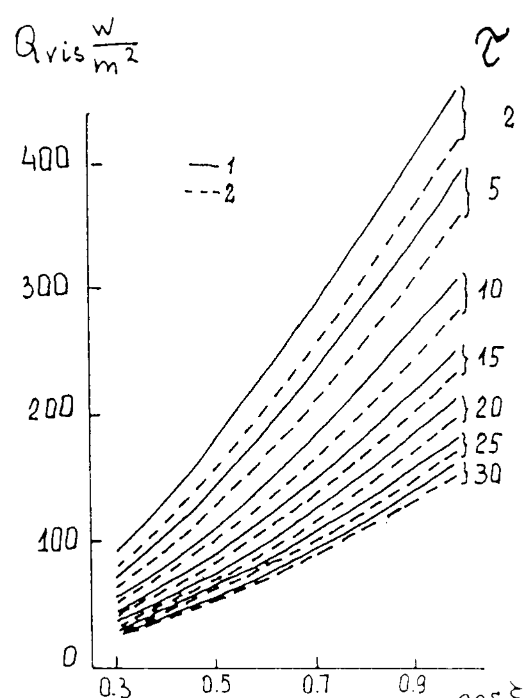

Fig. 1. Calculations of Q_{vis} for different τ, τ_{α,λ_o} and ξ; curve 1 – $\tau_{\alpha,\lambda_o} = 0.1$; 2 – $\tau_{\alpha,\lambda_o} = 0.9$. Albedo of the ocean 0.05; $r_{ef} = 9,8\mu$m, Sc cloud.

Then, comparing the measured and calculated Q_{vis} knowing ξ, one gets τ.

It is preferable to use Q_{vis} – to exclude absorption by liquid water and water vapour.

The results are given by Leontyeva (1992) for Atlantic ocean Sc clouds.

Figure 1 shows the influence of τ, ξ and τ_{a,λ_o} on Q_{vis} for Sc clouds with $r_{ef} = 9.8$ mkm.

The influences of aerosol – τ_{a,λ_o} is important only for small values of τ. If Q_{vis} is measured, τ can be obtained and the total Q can be calculated, taking into account the effective droplet radius for each cloud type, ground albedo, absorption by water vapour and cloud droplets.

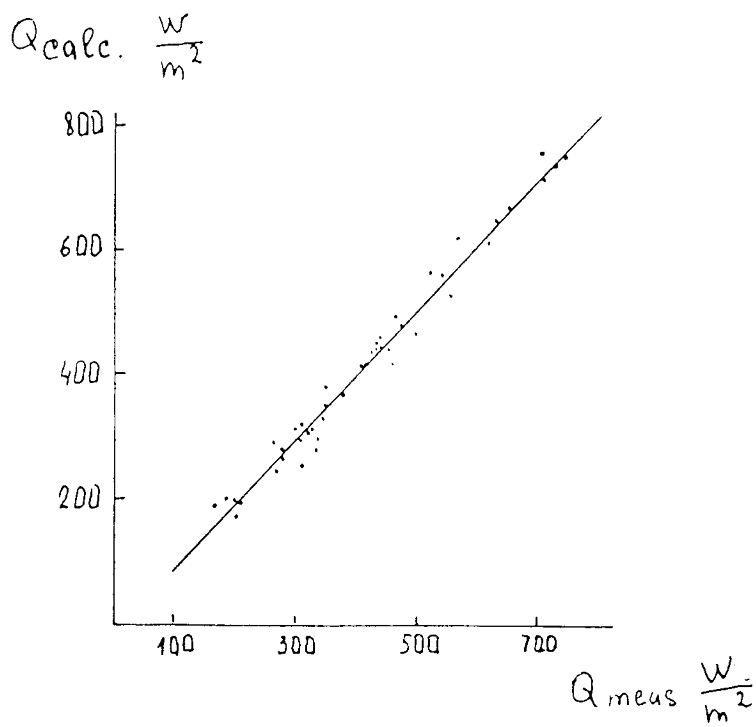

Fig. 2. Relation between total Q calculated and measured; Sc and As clouds.

Figure 2 gives the relation between total Q measured and calculated.

To diminish the influence of aerosols extinction, zenith angel of the sun and water vapour absorption, it is useful to consider the ratio C_Q of the total global radiation under cloudy conditions– Q_{cloud} to the same under the case of clear sky– Q_{clear}. The latter may be taken from time intervals when the sun is unshielded by clouds, or from mean data for clear sky with the same synoptic and other conditions.

Tarasova and Chubarova, 1994 calculated $C_Q = Q_{cloud}/Q_{clear}$ with taking into account absorption and scattering of light by cloud and aerosols particles, molecular scattering, absorption by water vapour and ozone.

The results for the global radiation were expressed in the following way: for

$$5 \leq \tau \leq 20 C_Q = (a + b\theta) \exp((c + d \cos \theta)\tau) \tag{5}$$

$$20 \leq \tau \leq 50 C_Q = \exp(-\tau/a(1 + \tau/b)^{0.5} \tag{6}$$

$$50 \leq \tau \leq < 100 C_Q = (a + b\theta)\tau^{(c+d\cos\theta)} \tag{7}$$

Here θ–the zenith angle of the sun in radians; constants a, b, c, d are given in Table 3 for different ranges of τ

Table 3. Constants in equation (5)–(7) and the error the approximations–$\Delta\tau$.

τ	a	b	c	d	$\Delta\tau$
5-20	0.961	-0.145	-0.053	-0.004	< 1
20-50	5.138	-1.073	-0.856	-0.01	< 1.7
50-100	9.477	-1.447	-1.043	-0.024	< 1.8

If C_Q is derived from measurements and compared with these formulas 15)–17), we get τ.

3. Determination of the cumulus cloud amounts on the bases of continuous recording of the total solar global radiation

The sensitivity of radiative fluxes to the presence of Cu clouds makes it possible to apply the following simple technique for determination their amount.

If we have continuous recording of the global solar radiation in the case of cumulus or other broken clouds, the cloud amount N may be calculated from the ratio:

$$N = 10 \sum_{i=1}^{m} \Delta t_i / \Delta t \tag{8}$$

where Δt_i–time intervals ($i = 1; \ldots$ m) when the record shows clouds (small value of Q), Δt– the whole time interval. Examples are given on Figure 3. Intervals t_i are calculated at the middle lines where $Q = 0.5(Q_{max} + Q_{min})$. The got values of N are slightly lesser than given by visual observations. That effect is clearly shown in Figure 4. By $N > 3$ the visual observations give bigger N than instrumental. The latter gives almost equal results, including the one used by our method (curve 4).

Figure 4 shows the middle distributions of cumulus cloud amounts in the center of the Equatorial Atlantic (1974, GATE expedition, ship "Academician Kurchatov"), evaluated by different ways (see Abranov *et al.*, 1976).

The visual observations overestimate the number of cases with big cloud amounts due to the perspective argumentation of clouds towards the horizon.

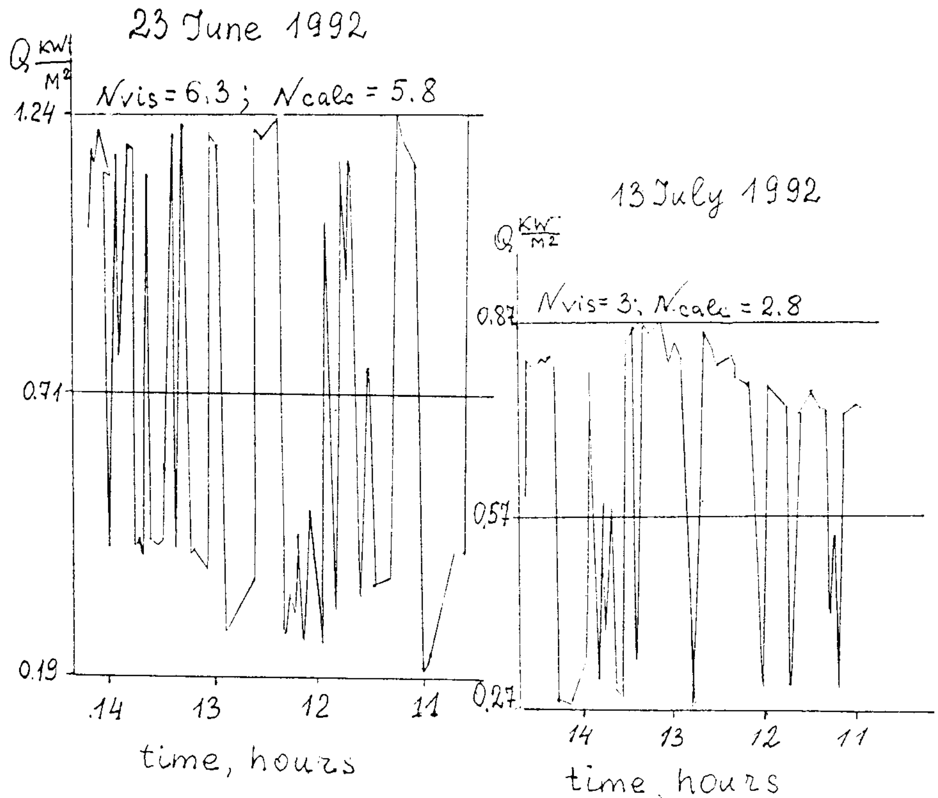

Fig. 3. Registrations of the global radiation Q and evaluations of the time intervals Δt_i: lines where $Q = 0.5(Q_{max} + Q_{min})$; N_{vis}– the results of visual observations; N_{cal}– calculations according to equation (8).

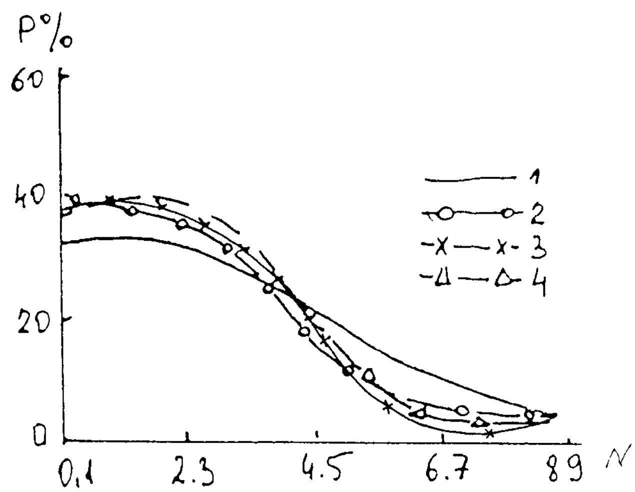

Fig. 4. Evaluation of the cloud amount by different techniques. P%– percentage of cases with $N = 0 - 1$; 2–3 etcetera. Curve 1– visual observations; 2 photography of the sky reflected from a half spherical mirror; 3– data of the optical sensor looking in the zenith; $\lambda = 0.7\mu$m, narrow angle of view; 4– data pyranometric measurements of global radiation.

4. Determination in case of unclouded atmosphere of the outgoing flux of the total thermal radiation – $F_{\uparrow}(\infty)$, if the corresponding downward flux at the ground $R_{\downarrow}(0)$ is given and vice versa (Galin *et al.*, 1991).

These fluxes are described by equations (Feigelson, 1973):

$$F_{\uparrow}(\infty) = B(0)D(0, Z^*) + \int_o^{z^*} B(Z')dD(Z^* - Z')/(dZ')dZ' \tag{9}$$

$$F_{\downarrow}(0) = \int_o^{z^*} B(Z')\frac{dD(0, Z')}{dZ'}dZ' \tag{10}$$

Here $Z^* = 20 - 30$ km – the upper level used in the calculations:

$B(Z) = \sigma T^4(Z)$; $D(Z)$ – the total transmission function in the layer $(0; Z')$ or $(Z'; Z^*)$ with $0 \leq Z' \leq Z^*$; $D(0) = 1$. Absorption by gases H_2O, CO_2, O_3 is taken into account.

Using the mean value theorem in (9), (10), we get expressions:

$$F_{\uparrow}(\infty) = B(0)D(Z^*) + B_1[1 - D(Z^*)] \tag{11}$$

$$F_{\downarrow}(0) = B_2[1 - D(Z^*)] \tag{12}$$

Then

$$\frac{B_1}{B_2} = \frac{F_{\uparrow}(\infty) - B(0)D(Z^*)}{F_{\downarrow}(0)} \tag{13}$$

If we know the ratio B_1/B_2, the ground temperature $T(0)$ and the whole content of H_2O vapour - W (taking the column density of O_3 and CO_2 as constants), we may evaluate $F_{\uparrow}(\infty)$ if we know $F_{\downarrow}(0)$ and vice versa.

Table 4 shows values of B_1/B_2 for standard models of the atmosphere. (Mc Clatchey *et al.*, 1972).

Table 4. Values of B_1/B_2 for standard atmospheres with corresponding total transmission function of the whole atmosphere column – $D(Z^*)$ and fluxes $F_{\downarrow}(0)$, $F_{\uparrow}(\infty)$.

Model		$T(0)K$	$W \frac{g}{cm^2}$	$D(Z^*)$	$F_{\downarrow}(0)\frac{w}{m^2}$	$F(\infty)\frac{w}{m^2}$	B_1/B_2
Tropics		300	3.8	0.114	397.8	294.4	0.608
Middle lat.	summer	294	2.7	0.157	348.9	286.6	0.631
Middle lat.	winter	272	0.77	0.287	218.8	235.9	0.669
Subarctic,	winter	257	0.37	0.327	166.7	202.6	0.728

Figure 5 shows the dependence of B_1/B_2 on the ground temperature.

$-T(0)°C$ and water vapour content W with corresponding $D(Z^*)$ for the region of the North-East Atlantic and the Mediterranean sea. The data of the Table 4 are also given.

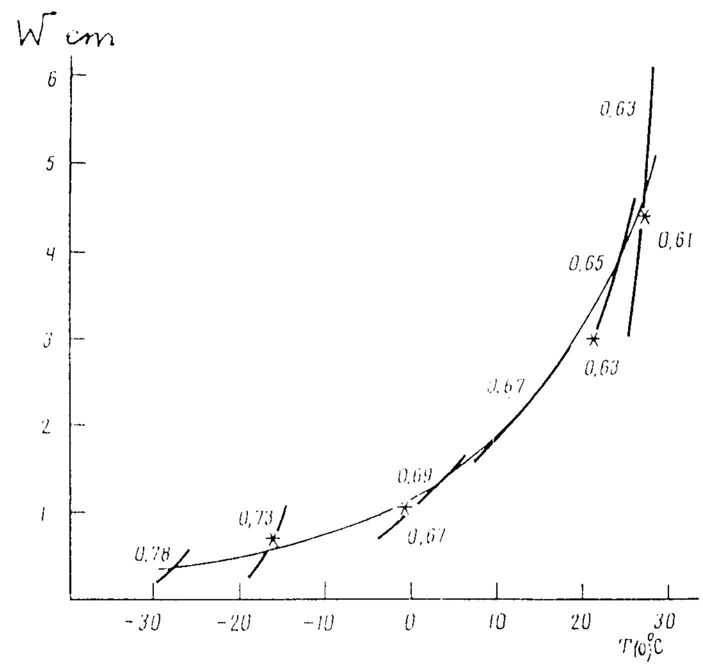

Fig. 5. The dependency of B_1/B_2 on $T(O)$ and W; stars– data of Table 4.

Another way of establishing the relation between the fluxes $F_\downarrow(0)$ and $F_\uparrow(\infty)$ is to make a set of calculations according to the equations (9), (10) for different temperature and water vapour distributions with height in a given region. Then to construct a regression equation. An example of it for the region mentioned above is given in Galinatae, 1991.

$$F_\downarrow(0) = 1.63[F_\uparrow(\infty) - B(0)/D(m^*)] - 4.33 \tag{14}$$

Till now we have considered the clear atmosphere.

If solid cloud and its top temperature Z_t are known, a well as the vapour content above the cloud, we can use a similar technique.

We can get $F_\downarrow(Z_t)$ if $F_\uparrow(\infty)$ is known and vice versa. That is only an idea; the corresponding calculations are not finished.

Conclusion

We think that the proposed ways to solve retrieval problems on the basis of standard based – ground actinometry are easy in application and fruitful. It is not unlikely that there are other ways of this kind.

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