Aerosol Contribution to Visible Skylight Polarization as Measured at Different Values of Underlying Surface Albedo

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Introduction

Correlation coefficients R between the polarization P of scattered light radiating by the day sky as well as twilight sky and aerosol optical depth τ_a, were analyzed in details at [1, 2] using the experimental data that had being measured till 1971. Most of the works on measurements of the day sky P at the Sun heights $h_{\odot}>0$ gives the $R\approx0.9$. That point had lead the authors of [1, 2] to conclusion that the P can be a sure indicator of optical properties of the atmosphere, especially about the τ_a.

However, as it is noted in [1, 2] some works contain a hardly explainable difference between the spectral dependence $P(\lambda)$ (known as the dispersion of the polarization; λ - light wavelength) and the R. It was supposed that the difference could be caused by the surface albedo A varying from one observational place to another as well as by insufficient experimental data. So, further measurements of the day sky P at the $h_{\odot}>0$ were not continued.

By now we have carried out a new investigation of the day sky P of the light irradiating from zenith in dependence on the positive $h_{\odot}$. In contrast to the other studies this one is realized at different values of albedo A (at presence of a snow cover as well as at its absence) both for reliable estimating the correlation coefficient R and for clarifying the role of the aerosol polarization in the variations of the $P(\lambda)$.

Instruments and Methodics

The experiment was carried out at Zvenigorod Research Station of IAP RAS from August 2003 to August 2004 using the instrumentation complex [3] including the original spectropolarimeter, a standard photometer and a standard actinometer. All the instruments have visual angle about 10° and are fully automated with a multi-channel A/D converter and a computer. Relative intensities of zenith day sky radiance I_λ were measured using the spectropolarimeter in the four spectral intervals that centered at the wavelengths λ_1=413 nm, λ_2=593 nm, λ_3=724 nm, λ_4=1005 nm. All the intervals have the half-bandwidth $\Delta\lambda\approx10$ nm. Below the λ index of appropriate values will be removed or be replaced with the number of the spectral interval.

The I_{1-4} are registered as rectangular pulses with duration about 5 sec. The full measuring cycle of all the I_{1-4} is about 30 sec (see the Figure 1). The spectropolarimeter is equipped with a rotating polaroid (rotation speed is 0.5 sec^{-1}) polarizing the visible light so the device can measure both the normal I_{max}

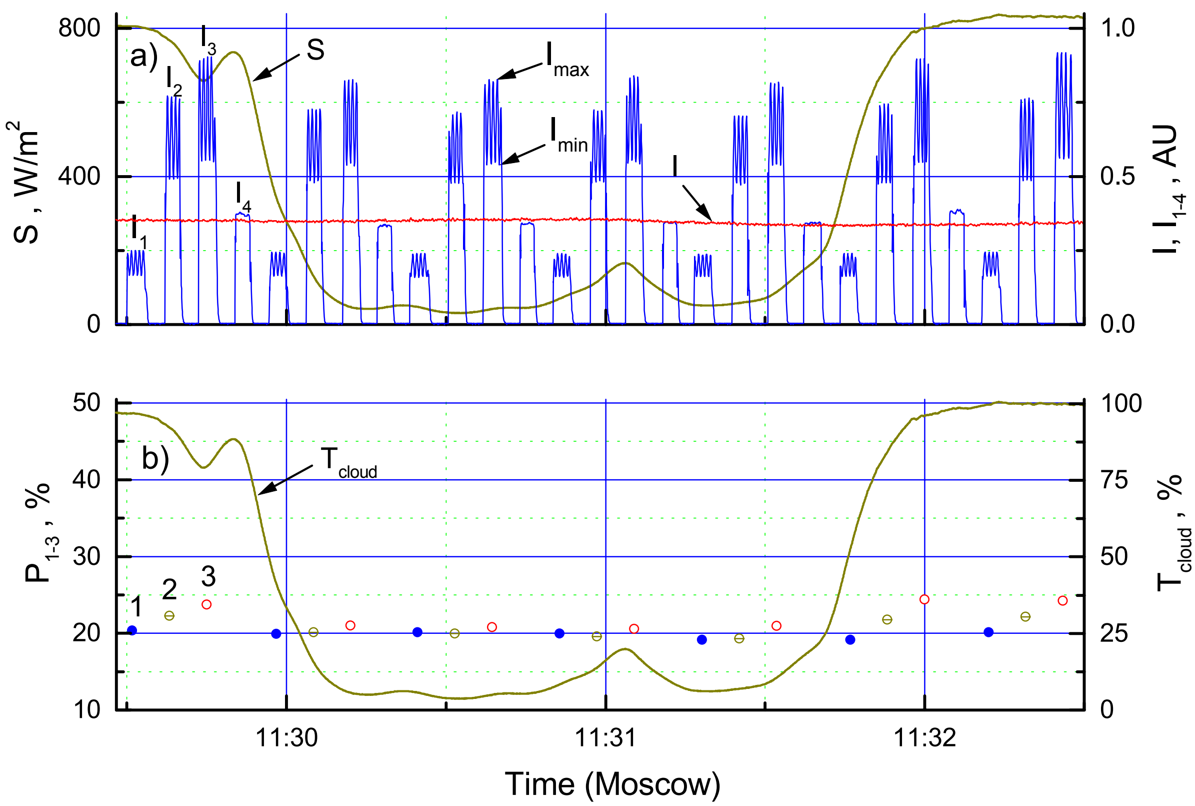

Figure 1a. An example of simultaneous scanning of all values of the sky radiance at four wavelength $I_{1\text{-}4}$, I, S at the $\tau_a \approx 0.16$ and at the $h_\odot$=39.6° at April 18, 2004 (Moscow time). **Figure 1b**. The cloud optical transparency T_{cloud} from total irradiance by pyrheliometer and the results of calculations of the appropriate polarization $P_{1\text{-}3}$. See also the text.

and the tangent I_{min} components of the light polarization three times per one measuring each of the I_{1-4} (see the Figure 1). Finally, the polarization P was calculated for the spectral intervals using the equation $P=(I_{max}-I_{min})/(I_{max}+I_{min})$ with accuracy about 3% for the λ_1 and about 1.5% for the $\lambda_{2,3}$.

The integral day sky irradiance I was measured at the wavelength range λ= 700÷2500 nm using the photometer and was applied to the control of the zenith sky haziness.

The actinometer was used to measure the direct solar radiance S (W/m^2). The values S were measured at the wavelength range 400-4000 nm and are corresponded to the effective wavelength λ_{ef}=550 nm. The actinometer was calibrated matching with the Eppley pyrheliometer that is standard device for measuring both the short wave and the long wave radiance at the BSRN network. Digital photo shots of zenith sky were being acquired simultaneously.

During more the one hundred measuring days a large amount of data were obtained for the values of τ_a in the range 0.04-0.4 and air temperatures t from –25°C to +25°C. The values of aerosol optical depth τa were estimated using the standard method taking into account temperatures t, a distance between the Sun and the Earth r and a value of water vapor content in the atmospheric column w, cm (courtesy by CAO Roshydromet).

The values of P are used for estimating the R using the equation $\tau* = -(1/m)\cdot \ln P/P^0 + \tau_{a,min}$, where m is the air mass, P^0 is the polarization at the highest observed air transparency at the $\tau a,min \approx 0.04$. Further, the dependencies $P0(h_\odot)$ are calculated at different values of albedo A (see the Table 1; A_1 corresponds to absence of a snow cover, A_2 relates to a snow cover after a rich snowfall). The τ∗ were estimated at $h_\odot$ in the range 15°÷40° with an inaccuracy about 3% for both the A_1 and the A_2.

Table 1. The polarization P^0 (at the lowest observed τa) vs. the Sun heights $h_\odot$ at the different surface albedoes: A_1 (snow cover is absent) and A_2 (snow cover is present after a rich snowfall).

	$h_\odot$	10°	15°	20°	25°	30°	40°	50°
A_1	$P^0 1$	67	61.1	54.8	48.6	42.5	30.1	18.6
	$P^0 2$	70.9	64.1	57.3	50.4	43.7	30.8	19.3
	$P^0 3$	69.8	62.8	55.7	48.6	41.6	28.8	17.4
A_2	$P^0 1$	58.7	51.7	45.0	38.4	33.0	-	-
	$P^0 2$	66.2	56.8	48.6	41.7	36.0	-	-
	$P^0 3$	66.2	56.8	48.8	42.2	36.5	-	-

Results

Using correlation diagrams τ∗ vs. τ_a that were plotted using the data of about 1000 all season realizations at clear sky the values $R_1=0.78$ and $R_2=R_3=0.96$ at the λ_{1-3} were found. The relatively low value R_1 is because of the Forbes effect. The empirical regression $\tau_a=0.83\cdot\tau*$ at the wavelength λ_2 =593 nm close to the λэф≈550 nm can be used retrieving the τ_a. For the operative purposes the values of τ_a can be retrieved using the values of τ∗ with accuracy about 10%. In that case there is no need to take into account the values of t, r and w.

If both τ_a and A are low (no snow cover) then the spectral dependence $P^0(\lambda)$ has practically neutral behavior. At the moderate as well as at the large τ_a the P grows with λ increasing. Because of that the variations of the polarization P were especially analyzed at conditions of shadowing the surface layer air with a cloud. If the Sun is at a gap between clouds the values of I as well as of I_{1-4} are large than if the surface layer is shadowed by a cloud because of decreasing the scattered radiation in the layer under the cloud (see Figures 1, 2).

An example of the scanning the values I, I_{1-4} and S are presented at Figure 1a; the values of P_{1-3} (expressed in %) and the cloud transparencies $T_{cloud}=(S_1/S_2)\cdot 100\%$ are given at the Figure 1b (see also Table 2). The polarization P grows as λ increases in agreement with numerical calculations of A.I. Ivanov, G.S. Livshits, V. Pavlov et al. (1968) [4]. Additionally, the results of continuous measurement of P_3 (obtained at August 3, 2004; $h_\odot=48.3°$, $\tau_a \approx 0.17$) are given at Figure 2. The values of P_3 are about 13.8±0.2% if the Sun shines at the surface layer air or 12.9±0.2% if the air layer is

Table 2. The polarization P vs. the light wavelengths λ_{1-3} at the different cloud transparencies T_{cloud}

T_{cloud} (%)	$P^0{}_1$	$P^0{}_2$	$P^0{}_3$
100	20.3 ± 0.3	22.5 ± 0.3	24.7 ± 0.2
8	20.0 ± 0.1	19.9 ± 0.3	20.8 ± 0.2
100	19.8 ± 0.6	22.0 ± 0.2	24.3 ± 0.1

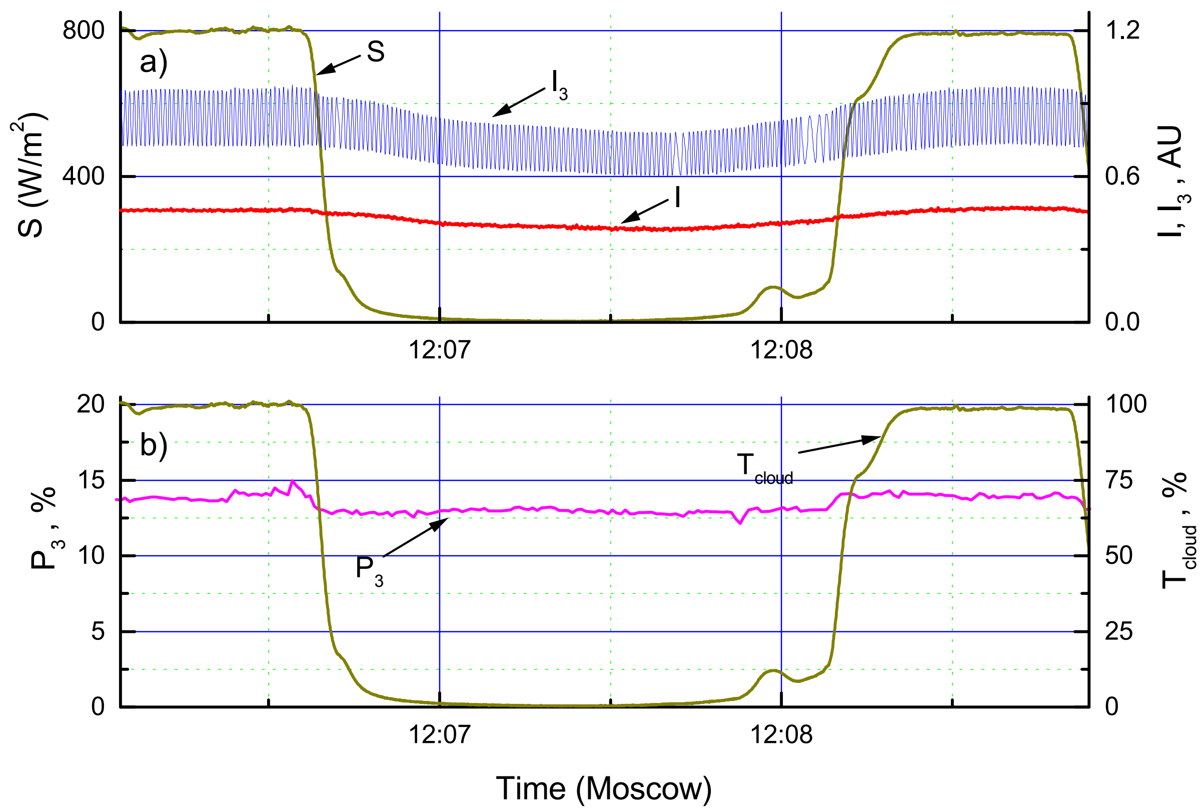

Figure 2a. An example of simultaneous scanning of all the I_3, I, S at August 03, 2004 (Moscow time).
Figure 2b. The results of the calculation of the appropriate P_3 and T_{cloud}. See also the text.

shadowed by a cloud. The Figures 1 and 2 show that the features of the P(λ) variations discussed in [1, 2] are forced not only by the different albedo A but by the light polarization on aerosol particles.

After a rich snowfall (A is now A_2) the values $P^0{}_1$ are markedly less then the $P^0{}_2$ and the $P^0{}_3$ (see the Table 1). That point is related not only to the spectral dependence $P^0(\lambda)$ at the presence of a snow cover but also to the aerosol polarization taking place even at the smallest value of τ_a.

Conclusions

1. Our method allows to make a real time estimation of the aerosol optical depths τa using the polarization of zenith day sky radiance at the Sun height $h_{\odot}>0$. The method runs also at the low surface albedo A being at absence of snow cover as well as at the high ones after a rich snowfall.

2. The polarization of a skylight caused by scattering on aerosol particles is discovered experimentally.

3. The features of the spectral dependence $P(\lambda)$ are not only defined by the differences of both the τa and the A but also caused by the aerosol polarization of a skylight.

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