OF POLARIZATION OF THE RADIATION EMANATING UPWARD FROM THE OCEAN SURFACE AS A FUNCTION OF HEIGHT

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The results of field measurements of the degree of polarization of the radiation emanating from the ocean surface are analyzed at different heights. The analysis was performed on the basis of approximate technique for calculating the polarization characteristics of the radiation in the "ocean—atmosphere" system. This technique takes into account the polarization of sky radiation, ocean waves, and partial polarization of radiation emanating from the water column. The influence of the observation height on the results of remote measuring the water turbidity with the use of polarization technique is studied.

The majority of studies performed up to date has indicated that the data on polarization characteristics of the radiation improved an information content of the passive methods for remote sensing of natural water in the visible range.[1-5] One of the important problems originating in the process of practical realization of corresponding technique is that of the distorting effect of the atmosphere, different aspects of which were examined in Refs. 2 and 6–11. The results of numerical calculations of the degree of polarization P of outgoing radiation in the "ocean-atmosphere" system were given in Refs. 6–8. The polarization characteristics of radiation at different heights were analyzed in Ref. 9 neglecting the roughness of the ocean surface. The semiempirical technique for the correction of the effect of the atmosphere in polarization measurements[10] is valid only for the observations in the plane of solar vertical (PSV). Reference 11 compares the results of computations obtained on the basis of the technique given in Ref. 10 with the results of experimental measurements of P at different heights H for the vertical angle of sight with respect to nadir $\theta = 45°$ and azimuth angle with respect to PSV $\varphi = 180°$. The height variation of the radiation emanating from the even surface $P(H)$ at the Brewster angle $\theta = 53°$ as a function of φ was studied in Ref. 2. In this paper the effect of different factors of the real "ocean—atmosphere" system (aerosol, roughness of the ocean surface, and polarization of the radiation emanating from the water column) on $P(H)$ is investigated.

The results of numerical simulation are compared with the experimentally obtained dependencies $P(H, \varphi)$. The influence of the observation height, at which the degree of polarization of upward radiation is measured, is studied as applied to the method of remote determination of water turbidity.[1]

Calculation method. A model of a plane stratified aerosol—molecular atmosphere with a horizontally homogeneous underlying surface, which models the ocean surface, is considered. The air—water boundary is assumed to be rough. The direction of observation for a given spectral range $\Delta\lambda$ is characterized by the nadir θ and azimuth φ angles and the optical thickness τ counted off the upper atmospheric boundary ($\tau = 0$). The polarized radiation field is described in terms of the Stokes parameters determined with respect to the meridian planes S_k, $k = 1, 2, 3$ and the degree of linear polarization $P = \sqrt{S_2^2 + S_3^2/S_1}$. The boundary conditions for the downward radiation for $\tau = 0$ are $S_k(0, \theta, \varphi) = \pi F_k \delta(\theta - \theta_0)\delta(\varphi)$ and $F_k = (F_0,0,0)$, where πF_0 is the solar constant, θ_0 is the zenith solar angle, and $\varphi_0 = 0°$ is the azimuth solar angle. For the upward radiation at the height of the underlying surface the condition of reflection from the rough surface and the water column is specified.[5] Owing to small values of the coefficient of diffuse reflection from the water column R (in the visible range $R \lesssim 10\%$), double reflections from the surface or water column were neglected. In this case the Stokes parameters for upward radiation may be given in the form

$$S_k = S_{\gamma,k} + S_{\rho,k},\ k = 1, 2, 3, \qquad (1)$$

where the first summand describe the photons which are multiply scattered in the atmosphere without interaction with the underlying surface and the second summand describes the contribution of the photons singly reflected from the ocean surface or water column and then scattered in the atmosphere. To determine $S_{\gamma,k}$, we used the Sobolev approximation, which states that the multiple scattering must be accounted for only in calculating the brightness S_1, where as the polarization characteristics are calculated in the single scattering approximation.[12]

The values $S_{\gamma,l}$ were found by numerical integration of the radiative transfer equation using the iteration method[2] in which we replaced the collision integral by singular squarings on unit sphere and invert the differential operator by integrating along the characteristics. The comparison with model calculations showed that the maximal error in calculating $S\gamma,_1$ was not greater than 3-5%.[14]

The Stokes parameters of singly scattered solar radiation $S^{(1)}_{\gamma,k}$ for arbitrary atmospheric stratification can be expressed in terms of the one—fold integrals

$$S^{(1)}_{\gamma,k}(\tau,\ \theta,\ \varphi) = \left(F_0/4\mu\right)e^{\tau/\mu} \times \int_{\tau}^{\tau_0} \Lambda(t)\exp\left(-t\left(\mu^{-1}+\ \mu_0^{-1}\right)\right)\gamma_{k1}\left(t,\ \mu,\ \varphi,\ \mu_0,\ 0\right)dt. \tag{2}$$

Here $\tau = \tau_a + \tau_R$ is the total optical thickness of aerosol (a) and molecular (R) atmosphere, $\mu = |\cos\theta|$, and $\Lambda(t)$ is a single scattering albedo. The normalized scattering matrix can be expressed in terms of the combination

$$\gamma_{km}(\tau,\ \theta,\ \varphi) = C_a(\tau)\ \gamma_{km,a}(\theta,\ \varphi) + \left(1 - C_a(\tau)\right)\ \gamma_{km,R}(\theta,\ \varphi) \tag{3}$$

of aerosol $\gamma_{km,a}$ and molecular (Rayleigh) $\gamma_{km,R}$ matrices including the rotation matrices. Here $C_a(\tau) = \beta_a(\tau)/(\beta_a(\tau) + \beta_R(\tau))$, where $\beta_a(\tau)$ and $\beta_R(\tau)$ are the aerosol and molecular volume scattering coefficient.

The data on the scattering matrix for polydisperse haze M, which models the marine aerosol,[1,3] were used as the matrix $\gamma_{km,R}$.

On the basis of the single scattering approximation, the expression for $S_{\rho,k}(\tau,\ \theta,\ \varphi)$ may be represented In the following form:

$$S_{\rho,k}(\tau,\ \theta,\ \varphi) = \left[T_{km}(\tau,\ \tau_0,\theta,\ \varphi) + T_0(\tau,\ \tau_0,\theta,\ \varphi)\delta_{km}\right]S_{\rho,m}(\tau_0),\quad k,\ m = 1,\ 2,\ 3, \tag{4}$$

where $S_{\rho,m}(\tau_0)$ are the Stokes parameters of upward radiation at the high of the underlying surface,

$$T_0(\tau,\ \tau_0,\theta,\ \varphi)\left[S_{\rho,k}(\tau_0)\right] = \exp\left[-(\tau_0-\tau)/\mu\right]\ S_{\rho,k}(\tau_0,\ \theta,\ \varphi)\ ,$$

and

$$T_{km}(\tau,\ \tau_0,\theta,\ \varphi)\left[S_{\rho,m}(\tau_0)\right] = (1/4\pi\mu)\ e^{\tau/\mu} \times \int_{\tau}^{\tau_0} \Lambda(t)\ e^{-t/\mu}\left\{\iint_{\Omega} \gamma_{km}(t,\ \theta,\ \varphi,\ \theta',\varphi') \times S_{\rho,m}(\tau_0,\ \theta',\varphi')\exp\left[-(\tau_0-\ t)/\mu'\right]d\omega'\right\}dt\ ,$$

where

$$d\omega' = \sin\theta' d\theta' d\varphi'\ ,$$

are the operators of straight and diffuse transmission of the atmospheric layer, and Ω is the upper hemisphere of directions. Repeated indices are assumed to be summed.

The expression for $S_{\rho,m}(\tau,\ \theta,\ \varphi)$ can be conveniently written as the sum of the Stokes parameters

$$S_{\rho,m} = S_{r,m} + S_{s,m} + S_{e,m}\ ,\quad m = 1,\ 2,\ 3, \tag{6}$$

of the straight solar radiation reflected from the rough surface $S_{r,m}$ the diffuse sky radiation $S_{s,m}$, and the radiation emanating from the water column $S_{e,m}$ (Ref. 5). The components $S_{s,m}$ and $S_{r,m}$ were calculated based on the model of randomly oriented surface elements, the wave slope distribution as a function of the wind speed V was described by the Cox—Munk function.[15]

In general the determination of the parameters of radiation emanating upward from the water column $S_{e,m}$ would call for solving the problem of the radiative transfer in the "ocean-atmosphere" system. Neglecting the double reflection pf photons from the ocean surface, however, we can model the water column by the effective underlying surface with the albedo equal to the spectral brightness coefficient $\rho(\lambda) = \pi\ I(\lambda)/E(\lambda)$, where $I(\lambda)$ is the spectral brightness of the upward radiation, $E(\lambda)$ is the spectral irradiance of a horizontal surface element from above immediately under the ocean surface which is determined using $S_{\gamma,k}(1)$ with account of refraction at the even boundary. The angular distribution of I was assumed to be isotropic. The angular distribution of polarization P_e of the radiation emanating from the water column was fitted by the one-parameter dependence typical for a fine fraction of terrigenous suspension, which determines backscattering in the ocean water.[16] The maximum value of the angular dependence

$P_e(\alpha_s) = P_e$ (90°), where α_s is the scattering angle of the straight solar beams in water column with refraction at the plane boundary taken into account, has proved to be variable parameter of the model that takes into account both microphysical properties of suspension and depolarization effects due to multiple scattering in ocean water.[5] The direction of the preferred polarization of the radiation emanating upward from the water column was assumed to be perpendicular to the scattering plane in the vicinity of the boundary.

Experimental studies. Measurements of P at different heights were performed above the Caspian Sea from the helicopter MI–8 under conditions of clear sky with the help of polarimeter described in Ref. 17. Some results were published in Ref. 11. Sighting of water surface was carried out at an angle $\theta = 45°$ with nadir. As shown in Refs. 1 and 5, the angle $\theta = 45°$ is close to optimal in determining the water quality by the polarimetric method: the measurements are of little effect for small values of θ in virtue of the small values of P and for $\theta > 45°$ in virtue of the decrease in the signal–to–noise ratio. The accuracy of measuring P is about 10-15%. The experimental results were obtained under the following conditions: zenith solar angle $\theta = 48–53°$, meteorological wind speed $V_m = 3$ m/s, sea roughness 2, visibility range 10 km, and clear water surface. The spectral brightness coefficient of water column $\rho_e(\lambda)$ and $P_e(90°)$ were not measured.

Calculation analysis. Let us consider the effect of the input parameters of the model (τ_a, V, ρ, $P_e(90°)$) on the dependence $P(H)$ for different values of φ. The calculations were performed at $\lambda = 450$ nm, where the effect of the atmosphere was maximum, and for zenith angles $\theta = 45°$ and $\theta_0 = 50°$.

The effect of τ_a and V on $P(H)$ for sighting in the plane of solar vertical $\varphi = 0°$ and 180°) is shown in Fig. 1. The height dependence $\tau_a(H)$ was assumed to be exponential: $\tau_a(H) = \tau_{a,0}\exp(-H/H_a)$, where $H_a = 1.2$ km. The large values of $P(H)$ for $\varphi = 0°$ may be accounted for by sighting "the solar track" since the angle of reflection of solar beams from specular surface elements is close to the Brewster angle (for water ≈ 53°). The aerosol is responsible for the depolarization of upward radiation due to the increase in atmospheric optical thickness and the decrease in the degree of polarization of atmospheric haze. The increase in the value of V in the zone of glint reflections produces strong depolarizing effect.

The dependence of P on V remains valid up to the upper atmospheric boundary ($P(V) = 72\%$ for $V = 3$ m/s and 61% for $V = 10$ m/s when $\tau_{a,0} = 0.4$ and $\rho = 1\%$), that agree with the data of Ref. 6. The influence of ρ and $P_1(90°)$ on $R(H)$ in the glint reflection zone is insignificant. A comparison of calculated and experimental data for $\varphi = 0°$ (Fig. 1) testifies to fairly well agreement for $\tau_{a,0} = 0.3–0.4$ (these values of $\tau_{a,0}$ agree with the data on meteorological visibility range $L \sim 10$ km).

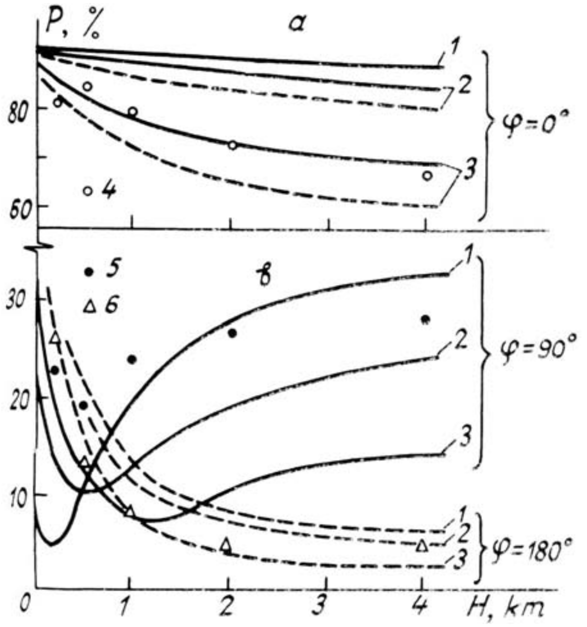

FIG. 1. Dependences $P(h)$ for $\theta = 45°$: 1), 2), and 3) are calculations at $\lambda = 450$ nm for $\theta_0 = 50°$, $\rho = 3\%$, $P_e(90°) = 0$, and $V = 5$ m/s; 1) Rayleigh atmosphere ($\tau_{a,0} = 0$), 2) and 3) composit e atmosphere, 2) $\tau_{a,0} = 0.13$ and 3) $\tau_{a,0} = 0.4$; a) $\varphi = 0°$ (dashed line is for $V = 10$ m/s); b) $\varphi = 90°$ and 180°; 4), 5), and 6) experimental results at $\lambda = 448$ nm for $\theta_0 = 48–53°$: 4) $\varphi = 0°$, 5) $\varphi = 90°$, and 6) $\varphi = 180°$.

When $\varphi = 180°$ atmospheric haze and the radiation emanating from the water column are practically uppolarized since the scattering angle of solar beams is close to 180°. For the values of P_0 at the high of the ocean surface the following expression is correct[5]:

$$P_0 = \frac{P_S}{1 + \xi_0(\lambda)}, \tag{7}$$

where P_S is the degree of polarization of the reflected sky radiation (of the component $S_{S,k}$), $\xi_0 = S_{1,1}/S_{S,1}$ is the ratio of brightnesses of the reflected radiation.[2]

The parameter $\xi_0(\lambda)$ is proportional to $\rho(\lambda)$, therefore P_0 decreases with increase of ρ when the conditions of the irradiance remains unchanged. The dependence $P_0(\rho)$ forms the basis of the polarization technique for the remote determination of water turbidity.[1,5] Numerical calculations demonstrate the dependence $P_0(\rho)$ to be very strong by virtue of the high values of P_S (P_S changes insignificantly from 90 to 82% as V increases from 3 to 10 m/s). For example, for $\tau_{a,0} = 0.4$ $P(\rho = 0.5\%) = 55\%$ and $P(\rho = 5\%) = 24\%$. The dependence $P(\rho)$ becomes weak, however, with increase of H by virtue of the depolarizing effect of atmospheric haze. Thus, under the same conditions of observations at $H = 500$ m $P(\rho = 0.5\%) = 14\%$ and $P(\rho = 5\%) = 11\%$.

The comparison with the experimental data enables us to estimate the value of ρ(450 nm) at the moment of measurement since for $\varphi = 180°$ the dependence of the degree of polarization $P_e(90°)$ on $\tau_{a,0}$ is weak.

As it follows from Fig. 1, the best agreement between the experimental and calculated results is achieved for p = 450 and is about 2-3%. For the values of φ close to 90° and 270° the dependence $P(H)$ is of substantially nonmonotonic character: for $H = H_0$ the well–pronounced minimum is observed (Fig. 1). An analysis performed in Ref. 2 showed that the appearance of the minimum is caused by the difference in directions of depolarization of the components $S_{\rho,k}$ and $S_{\gamma,k}$ (1). For $H < H_0$, $S_{\rho,k}$ prevails in the upward radiation, its direction of polarization being horizontal $S_2 \sim S_{\rho,2} < 0$). For $H > H_0$, atmospheric haze prevails. For $\theta_{0\sim} > 30°$ and $\theta_{0\sim} > 40°$ the direction of polarization of haze is close to the vertical plane $S_2 \sim S_{\rho,2} > 0$). That is why an increase in H forces the parameter $S_2(H)$ to increase rapidly, and at $H = H_0 S_2$ reduces to zero ($P(H)$ reaches its minimum). The presence of the aerosol manifests itself in increasing the height of the minimum H_0 and decreasing the values of $P(H)$ for $H \gtrsim 1$ km. These effects can be explained by the decrease of the degree of polarization of atmospheric haze and in particular, by a slower rate of growth of the parameter $S_2(H)$ in comparison with the Rayleigh atmosphere. The experiments prove that for $\theta_0 = 48°–53°$ the minimum value of $P \approx 20\%$ is observed at $H_0 \approx 500$ m. The calculated curves for $\tau_{a,0} = 0.3–0.4$ give the underestimated values of P in addition the differences for $H \gtrsim 500$ m exceeds the measurement errors. The calculated values of H_0 are also much greater ($H_0 = 1–1.2$ km). One of the possible reasons accounting for these disagreement may be the nonmonotonic character of the height dependence of aerosol particle number density. Let us examine the effect of this factor on $P(H)$ for $\tau_{a,0} = 0.4$ for three models of vertical aerosol distribution:

1) homogeneous layer located at heights varying from 1.5 to 3 km;

2) exponential dependence: $\tau_a(H) = \tau_{a,0} \exp(-H/H_a)$; where $H_a = 1.2$ km;

3) homogeneous layer located at $H \lesssim 500$ m.

As can be seen from Fig. 2, for $\varphi = 0°$ depolarization of the upward radiation takes place in the aerosol layer. For the Rayleigh atmosphere the dependence $P(H)$ is weak. On the upper atmospheric boundary P is determined by the value of $\tau_{a,0}$, and is practically independent of the type of vertical aerosol distribution.

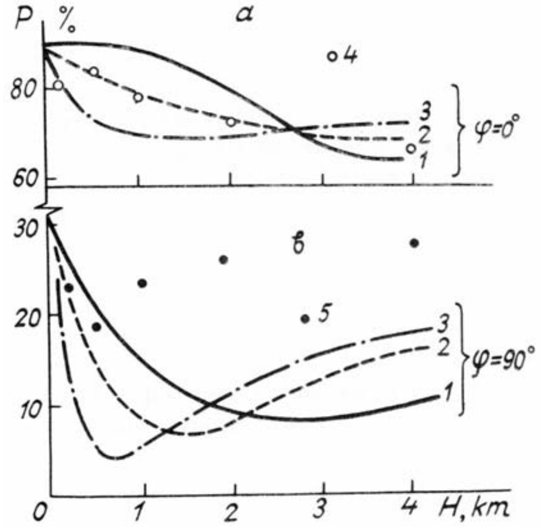

FIG. 2. Calculated dependences $P(H)$ at $\lambda = 450$ nm for $\theta_0 = 50°$, $\theta = 45°$, $\tau_{a,0} = 0.4$, $P_e(90°) = 0$, and $V = 5$ m/s for three models of vertical aerosol distribution:
1) homogeneous layer located at heights varying from 1.5 to 3 km;
2) exponential dependence, $H_a = 1.2$ km;
3) homogeneous layer for $H \le 500$ m;
4) and 5) experimental data at $\lambda = 448$ nm for $\theta_0 = 48–53°$: a) $\varphi = 0°$, b) $\varphi = 90°$.

For $\varphi = 90°$ the height of the minimum H_0 increases with the height of aerosol layer and $P(H)$ also increases. The minimum of $P(H)$ dependence becomes less pronounced for larger thicknesses of the layer. In general, the comparison with the experimental data (Fig. 2) testifies to the fact that the exponential model of vertical aerosol distribution is preferable possibly for smaller values of $H_a < 1$ km. The overestimated values of $P(H_0)$ as compared with the results of calculation can be caused by the partial polarization of radiation emanating from the water column.

The analysis performed showed that the effect of input parameters of the model on the dependence $P(H)$ manifests itself differently depending on the observation azimuth φ. This enables us to make comparison with the experimental data, when supplemental hydrooptical and actinometric data are deficient, by means of estimating the unknown parameters $\tau_{a,0}$, $P_e(90°)$, and $\rho(\lambda)$ following the technique shown above. The results of comparison performed at two different observation heights $H = 200$ m and $H = 1$ km are shown in Fig. 3.

The calculated and experimental data are given in the form of polar diagrams showing the variations in P depending on the angle φ for fixed values of θ and θ_0. The comparison shows that the angular distribution of polarization of upward radiation for all the values φ and H can be correctly described by the numerical model.

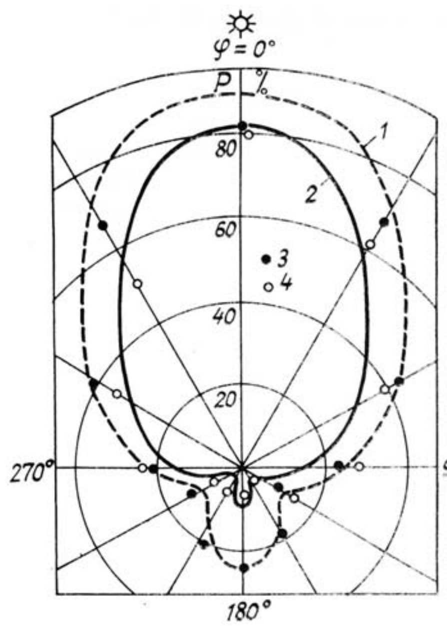

FIG. 3. Azimuthal dependences $P(\varphi)$ at H = 200 m and H = 1 km; 1) and 2) numerical calculations at λ = 450 nm, *for* $\theta_0 = 50°$, $\theta = 45°$, $\tau_{a,0} = 0.3$, $\rho = 2\%$, V = 10 m/s, *and* $P_e(90°) = 0.6$; *3) and 4) experimental results at* λ = 448 nm *for* $\theta_0 = 52°–53°$: *1) and* 3) H = 200 m *and 2) and 4)* H = 1 km.

Let us consider the problems of optimizing the remote polarization method of monitoring the quality of water based on the model described above.

The problem of the possibility of determining water turbidity from the data of measurements of P is of particular importance. A correlation dependence is found from the experimental data analysis[1] which shows that for a wide region of significant parameters any increase in the suspension concentration c_s results in the decrease of P_0. The tendency toward decrease of P_0 with increase of c_s can be explained by the increase in the brightness of weakly polarized radiation emanating from the column.[3,5] For the unpolarized underwater radiation ($P_e(90°) = 0$) formula (7) is correct for all the directions of sight. The value of $\rho(\lambda)$ increases with the concentration of terrigenous suspension c_s for the spectral range $\lambda \geq 550$ nm. For this reason the increase of the parameter $\xi_0(\lambda)$ Eq. (7)) and the decrease of P_0 are observed under constant conditions of irradiance. An attempt was undertaken in Ref. 5 to determine theoretically the optimum angles of sight θ and φ in the technique described in Ref. 1 for the case of measurements performed onboard the ship. For this purpose under constant conditions of irradiance it is convenient to introduce polarization contrast of the dependence $P(\rho)$

$$\eta_\rho(\theta, \varphi, P_e) = [P(\rho_1) - P(\rho_2)]/\delta P, \tag{8}$$

where $\delta P = 5\%$ is the absolute error in measuring P with the available equipment.[1]

Investigation of the angular dependences η_r at the height of the ocean surface for a wide range of variations of the model parameters V, θ_0, λ, and $P_e(90°)$ performed in Ref. 5 showed the measurements for $\theta \lesssim 30°$ to be of little effect by virtue of small values of P. For $\theta > 40°$ two zones of maximum contrast are observed for if $\varphi \approx 30$-$60°$ and $\varphi \gtrsim 150°$. Small values of P and strong effect of the conditions of irradiation make measurements difficult, that is why the angles $30° \lesssim \varphi \lesssim 53°$ and $40° \lesssim \theta \lesssim 53°$ were recommended in the technique described in Ref. 1. In this paper the angular dependences η_ρ are investigated at different heights. Figure 4 shows isolines of the values $\eta_\rho(\theta, \varphi)$ = const in polar coordinates in which the radius is set by the angle θ, and the polar angle is set by φ at H = 200 m and 2 km. The numerical calculation was performed at λ = 645 nm for $\tau_{a,0} = 0.1$ and $\rho_1 = 0.5\%$ and $\rho_1 = 2\%$, which are typical for the Caspian Sea.

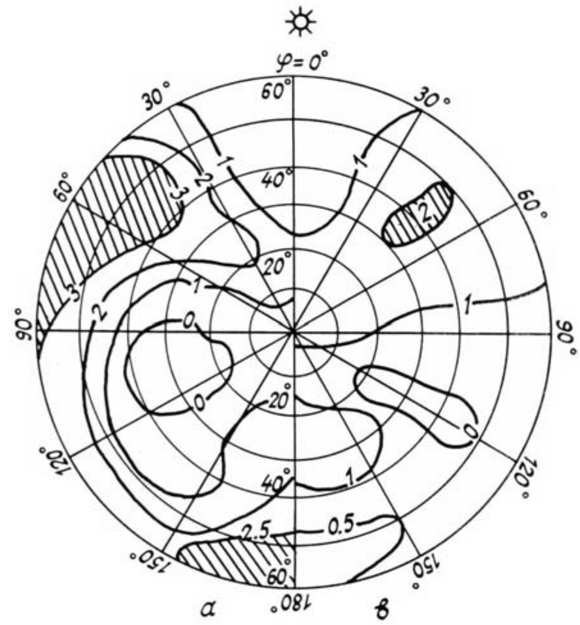

FIG. 4. Contrasts $\eta_\rho(\theta,\varphi)$. Numerical calculations at λ = 645 nm $\tau_{a,0} = 0.1$, $P_e(90°) = 0.2$, V = 5 m/s, $\rho_1 = 0.5\%$ *and* $\rho_2 = 2\%$: *a)* H = 200 m *and b)* H = 2 km.

The analysis of Fig. 4 showed that the contrasts η_ρ within the zone of optimal directions of sight $\theta = 45°$ and $\varphi = 45°$ remain high even at $H \approx 2$ km. Conversely, for $\varphi \gtrsim 150°$ the effect of the atmosphere manifests itself more strongly and these angles cannot be recommended for remote measurements.

The dependence of contrast $\eta_\rho(H)$ on atmospheric turbidity for optimal direction of sight ($\theta = 45°$ and $\varphi = 45°$) is of significance. Our data showed that the values of η_ρ for all the heights rapidly decrease with the increase of $\tau_{a,0}$. Thus, at

$\lambda = 645$ nm for $\tau_{a,0} = 0.3$ the dependence $P(\rho)$ is noticeable only for $H \gtrsim 1.5$ km. For smaller values of λ the effect of the atmosphere increases. Thus, at $\lambda = 450$ nm the maximal heights of observing the dependence $P(\rho)$ do not exceed 1 km.

CONCLUSIONS.

1. The model proposed in this paper adequately describes the height and azimuthal dependence of the degree of polarization of the radiation emanating upward ocean surface and the effect of the main factors of the real "ocean-atmosphere" system (τ_a (H), V, and ρ).
2. The range of optimal directions of sight for the polarization technique of remote determining water turbidity as a function of height is established.

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