EFFECT OF REGIONAL CHARACTERISTICS OF THE ATMOSPHERE ON THE ACCURACY IN DETERMINING THE TEMPERATURE OF THE OCEAN SURFACE FROM IR MEASUREMENTS FROM SPACE AT TWO ANGLES

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The errors in determining the temperature of the ocean surface (TOS) from remote measurements of IR emission at two angles relative to the vertical owing to the statistical variability of the altitude profiles of the atmospheric temperature and humidity are analyzed based on model calculations. The optimal coefficients and the corresponding errors in determining the TOS in the linear algorithm of atmospheric correction, demonstrating that they are highly variable, are determined in the approximation of an absolutely black surface and a cloud- and aerosol-free atmosphere for different regions in the northern hemisphere. The effect of errors in the radiation measurements is studied.

One of the promising method for the remote determination of the temperature of the ocean surface (TOS) is based on measurements of the infrared radiation at different angles relative to the vertical. A two-angle measurement scheme was first studied in order to make corrections for absorption of radiation by water vapor in the atmosphere.[1,2] Later aerosols and the dependence of the emissivity of the sea surface on the sea state were included among the factors taken into account and measurements at three and more angles became of interest.[3–6]

Infrared radiation in the "ocean-atmosphere" system (SOA) in the transmission windows depends on a number of factors including the vertical profiles of the temperature and humidity, so that the best approach to the analysis of the atmospheric effects is to use the general principles of the solution of the inverse problems of thermal sounding,[7] based on the use of statistical data on the variability of the atmosphere. However this approach is not widely employed for investigating the problem of determining the TOS. For spectral multichannel measurements it is implemented, for example, in Ref. 8. In this paper the two-angle method of atmospheric correction is studied based on the same assumptions. In so doing, in order to separate most clearly the effects associated with the effect of atmospheric temperature and water-vapor profiles, other interfering factors are not included in the radiation model employed.

For the ocean region selected approximate values of the TOS $T^{(0)}$ and vertical profiles of the atmospheric parameters can be given at each moment in time. Using a finite-dimensional approximation for the atmospheric profiles, we denote by $\vec{x}^{(0)}$ the approximate values of the temperature and humidity at N altitudes (here $\vec{x}^{(0)}$ is a $2N$-dimenslonal vector). Assuming that at an arbitrary altitude the values of these parameters are determined uniquely by a fixed interpolation from the starting levels, the measured radiation temperatures $\vec{T}_r$ can be assumed to be a function of the quantities T and $\vec{x}$: $\vec{T}_r = \vec{T}_r(T, \vec{x})$. According to the stated goal of this work $\vec{T}_r$ is assumed to be a two-dimensional vector, whose components describe the measurements at two angles. The real values of the TOS T and the vector $\vec{x}$, describing the state of the atmosphere, differ from the reference values and are given by $T = T^{(0)} + \Delta T$ and $\vec{x} = \vec{x}^{(0)} + \Delta\vec{x}$. The corresponding vector of measurements is $\vec{T}_r = \vec{T}_r^{(0)} + \Delta\vec{T}_p$.

Approximate calculations showed that for real variations of ΔT and $\Delta\vec{x}$ a local linear approximation of the function $\vec{T}_r = \vec{T}_r(T, x)$ was valid to a high degree of accuracy:

$$\Delta\vec{T}_r = \frac{\partial\vec{T}_r}{\partial T}\Delta T + \vec{a}; \quad \vec{a} = \sum_{1=1}^{2N} \frac{\partial\vec{T}_r}{\partial x_1}\Delta x_1, \tag{1}$$

where a is the two-dimensional vector with the components a_1 and a_2, and Δx_1 is the i-th component of the vector l$\Delta\vec{x}$.

This relation can be regarded as a system of linear equations, one unknown in which is ΔT. For this reason it is natural to express the solution of this system in the linear form $\Delta\hat{T} = (\vec{\alpha}, \vec{T}_p)$ or in the more general form $\hat{T} = \alpha_0 + (\vec{\alpha}, \vec{T}_p)$ (the parentheses indicate a scalar product); $\Delta\hat{T}$ and $\hat{T}$ denote the results of the determination of the TOS from remote sensing data and $T = T^{(0)} + \Delta T$. The error In determining the TOS from these formulas is:

$$\delta T = \left[\alpha_0 + \left(\vec{\alpha}, \vec{T}_r^{(0)}\right) - T^{(0)}\right] + \left[\left(\vec{\alpha}, \frac{\partial \vec{T}_r}{\partial T}\right) - 1\right]\Delta T + (\vec{\alpha}, \vec{a}). \quad (2)$$

It can differ from zero owing to the fact that the number of unknowns (ΔT and Δx_i) in the system of equations (1) Is greater than the number of equations as well as because of the fact that the coefficients α_0 and α (α is a vector whose components are α_1 and α_2) are probably not known adequately.

The dependence of the radiation temperature of the SOA on the TOS and the parameters of the atmosphere Is nonlinear, so that the derivatives (ΔT and Δx_i) entering the expressions (1) and (2), like also the quantities $\vec{T}_r^{(0)}$, should depend on the reference states of the TOS $\vec{T}^{(0)}$ and the atmosphere $\vec{x}^{(0)}$. Therefore the optimal values of α_0 and $\vec{\alpha}$ it which give the smallest errors in determining the TOS must be determined, generally speaking, separately for each reference state of the SOA. Analysis of this question for the example of the two-angle method is the basic goal of this work.

By definition only quantities which are constant for the given region enter the first term in the expression for δT. Hence in the theoretical analysis of the effects associated with the variability of the atmosphere this term may be assumed to be zero. For this the condition $\alpha_0 = T^{(0)} - (\vec{\alpha}, \vec{T}_r^{(0)})$ must hold. For a concrete value of α_0 the greatest difficulties under real conditions could be associated with setting the value of $\vec{T}_r^{(0)}$, for example, owing to the fact that a large number of undetermined factors must be taken into account in the model calculations of the absolute values of the radiation temperatures. In practice the best method for giving α_0 can apparently be based on direct referencing of the remote measurements at separate calibration points, for which simultaneous measurements of $\vec{T}_r$ and the TOS are available. Since the quantities $\vec{T}_r^{(0)}$ and α_0 enter the expression for δT only in the first term the problem of giving α_0 does not arise in the theoretical analysis of the other terms of the problem. For this reason a detailed analysis of this question need not be made in this paper.

The optimal vectors $\vec{\alpha}$ for each region must be determined by taking into account the second and third terms in (2). If it is additionally required that $\vec{\alpha}$ satisfy the condition $\left(\vec{\alpha}, \frac{\partial \vec{T}_r}{\partial T}\right) = 1$, then the quantity δT will be determined by only the third term. The convenience of giving $\vec{\alpha}$ in this manner lies in the fact that δT will not depend on ΔT. In this case the components of the vector $\vec{\alpha}$ are related by the formula $\alpha_2 = (1 - \alpha_1\tau_1)\tau_2^{-1}$ and the expression for δT finally assumes the following form:

$$\delta T = \alpha_1 a_1 + (1 - \alpha_1\tau_1)\tau_2^{-1} a_2. \quad (3)$$

Here the fact that $\frac{\partial \vec{T}_r}{\partial T} = \vec{\tau}$; was taken into account; the transmission of the atmosphere for the two viewing angles are denoted by τ_1 and τ_2.

One can see from formula (3) that in the case when a strong functional relation exists between α_1 and α_2 it would be possible to give α_1 so as to make δT equal zero. But since a_1 and a_2 depend on $2N$ variable quantities Δx_i, such a relation could be realized only if all vectors $\frac{\partial \vec{T}_r}{\partial x}$ are linear with one another. Numerical calculations show (see below) that this condition, strictly speaking, does not hold, and for this reason one can only hope that sufficiently strong statistical correlations exist between Δx_i (and hence between a_1 and a_2 also). Physically the assumption that strong correlations exist between the parameters of the atmosphere at different altitudes means that the temperature and humidity profiles vary significantly as a whole and in correlation with one another.

We shall determine the optimal values of the coefficients $\vec{\alpha}$ and the corresponding values of the error in determining the TOS by minimizing the variance of the latter. Without specifying for the time being the ensemble of states of the SOA we shall write out the basic relations under the assumption that $T^{(0)}$ and $\vec{x}^{(0)}$ are the mathematical expectations of the quantities T and $\vec{x}$. We denote by σ_1 and σ_2 the standard deviations of a_1 and a_2 and by ρ the correlation coefficient between them. By definition the mathematical expectation of the quantities Δx_i is equal to zero, so that the mathematical expectation of a_1 and a_2 are also equal to zero. The expression for the standard deviation of δT assumes, using (3), the form

$$\sigma^2 = \alpha_1^2\sigma_1^2 + 2\alpha_1 \frac{1 - \alpha_1\tau^1}{\tau_2}\sigma_1\sigma_2\rho + \left[\frac{1 - \alpha_1\tau_1}{\tau_2}\right]^2 \sigma_2^2. \quad (4)$$

Thus far we have omitted from the factors determining δT and σ the errors in the measurements of the radiation temperature. If these errors sire assumed to be random and Independent and have zero mean values and the same standard deviation σ_n for each channel, then it is sufficient to include the term $(\alpha_1^2 + \alpha_2^2)\sigma_n^2$ in expression (4) in order to take them into account. The optimal values of α_1, α_2 and σ with $\sigma_n \neq 0$ will be determined by the following relations:

$$\alpha_1 = \frac{1}{\tau_1}\cdot\frac{\nu-\mu\rho}{1-2\mu\rho+\nu};\quad \alpha_2 = \frac{1}{\tau_2}\cdot\frac{1-\mu\rho}{1-2\mu\rho+\nu};$$

$$\sigma = \frac{(\sigma_1^2+\sigma_1^2)^{1/2}}{\tau_1}\cdot\left[\frac{\nu-\mu^2\rho^2}{1-2\mu\rho+\nu}\right]^{1/2}, \qquad (5)$$

where

$$\nu = \frac{\tau_1^2}{\tau_2^2}\cdot\frac{\sigma_2^2+\sigma_n^2}{\sigma_1^2+\sigma_n^2};\quad \mu = \frac{\sigma_1\sigma_2\tau_1}{(\sigma_1^2+\sigma_n^2)\tau_2}.$$

The relations (5) were obtained from the condition $\dfrac{\partial\sigma^2}{\partial\alpha_1} = 0$.

We denote by H the matrix constructed from the column vectors $\dfrac{\partial\vec{T}_r}{\partial x_i}$ ($i = 1, \ldots, 2N$), by H^T the transposition of H; and, by G the matrix of covariations of the atmospheric parameters. The quantities σ_1, σ_2, and ρ can be determined in terms of the matrix Φ given by

$$\Phi = HGH^T;\quad \Phi = \begin{Bmatrix}\sigma_1^2 & \sigma_1\sigma_2\rho \\ \sigma_1\sigma_2\rho & \sigma_2^2\end{Bmatrix}. \qquad (6)$$

To obtain numerical it is necessary to choose a radiation model of the SOA for the chosen spectral interval as well as the values of $T^{(0)}$, $\vec{x}^{(0)}$, and G. The latter were taken from Ref. 9; in addition, in all cases $T^{(0)}$ was assumed to be equal to the temperature of the bottom layer of the atmosphere. We note that the data in Ref. 9 were obtained by averaging over a period of one month for summer and winter separately and in tropical regions for spring and fall also. In case of monthly averaging over the indicated regions the condition of statistical stability can apparently be regarded as satisfied.

The derivatives $\dfrac{\partial\vec{T}_r}{\partial T} \equiv \vec{\tau}$ and $\dfrac{\partial\vec{T}_r}{\partial x_i}$ were calculated for the spectral interval 900–920 cm^{-1}, and absorption by water vapor only was taken into account. The transmission is calculated as the product of the continuum and selective components. The continuum absorption was given according to Ref. 10 and the selective absorption was given according Goody's band model using the Curtis-Godson method for inhomogeneous atmospheres.[11] The generalized parameters of the lines required for the calculations were taken from Ref. 12 where they are presented for three temperature T_0 (220, 260, and 300°K) and the atmospheric pressure P_0. The generalized parameters of the lines are Interpolated to the running values of T and P with the help of the relations[10]

$$S = S_0\left(\frac{T_0}{T}\right)^{3/2}\exp\left[-\frac{E}{k}\left(\frac{1}{T}-\frac{1}{T_0}\right)\right];$$

$$\alpha = \alpha_0\left(\frac{P}{P_0}\right)\left(\frac{T_0}{T}\right)^{1/2},$$

where S and α are the intensity and half-width of the lines.

The transfer equation was integrated on an ES-1033 computer with double precision by the trapezoidal method up to an altitude of 10 km with a uniform step of 0.1 km. Linear interpolation of the profiles of the meteorological parameters at the integration nodes was employed for the temperature and exponential interpolation was employed for the humidity. Referencing of the pressure levels[9] to the altitude was performed using five standard atmospheres.[13] The surface was assumed to be absolutely black.

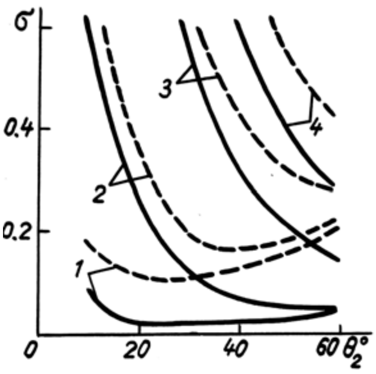

FIG. 1. The angular dependence of the errors in determining the TOS for two regions with different values of σ_n; the broken lines pertain to the region 4.6 (fell); the solid lines perteln to the region 3.1 (summer); σ_n = 0 (1), 0.01 (2), 0.05 (3), and 0.1°K (4).

The main computational results are presented In Figs. 1–4. Figure 1 shows the dependence of the optimal values of σ on the angle of observation θ for two characteristic atmospheric conditions – the region 4.6 (fall) according to the classification of Ref. 9 includes the Caribbean Sea, the Gulf of Mexico, and the adjacent part of the Pacific Ocean and the region 3.1 (summer) includes the eastern subtropical sector in the north Atlantic. Some numerical data for the same regions are presented In Table I. The angle θ indicates the direction of observation in the second channel (the angle is measured relative to the zenith at the point where the observation ray Intersects the surface); the first channel In all cases corresponds to observation at the nadir.

TABLE I.

Effect of the regional peculiarities of the atmosphere

Region	Season	Q, g/cm^2	σ, K	θ, deg	α_1	α_2	$\sqrt{\alpha_1^2+\alpha_2^2}$	σ	$\sigma_n\sqrt{\alpha_1^2+\alpha_2^2}$
4.6	fall	4.8	0	30	11.23	-10.17	15.15	0.13	—
				60	3.12	-2.02	3.72	0.22	—
			0.1	30	9.28	-7.99	12.25	1.37	1.23
				60	3.11	-2.00	3.70	0.43	0.37
3.1	summer	2.7	0	30	9.38	-8.38	12.58	0.02	—
				60	2.49	-1.48	2.90	0.05	—
			0.1	30	4.82	-3.60	6.02	0.87	0.60
				60	2.44	-1.42	2.82	0.29	0.28
2.6	winter	0.3	0	30	8.00	-7.00	10.63	0.00	—
				60	2.13	-1.13	2.41	0.00	—
			0.1	30	0.54	0.47	0.72	0.10	0.07
				60	0.76	0.26	0.80	0.12	0.08

The character of the dependence of σ on θ is largely determined by the value of σ_n. For $\sigma_n = 0$ the optimal angle $\theta = 20$–$30°$, but in the range $\theta = 20$–$60°$ the changes in or are small. For the most realistic level of error in the measurements $\sigma = 0.1$°K the dependence of or on в is stronger owing to the fact that as θ increases the quantity $|\alpha| = (\alpha_1^2 + \alpha_2^2)^{1/2}$ decreases substantially. This results in a sharp decrease in σ right down to $\theta = 60°$ (higher values were not studied).

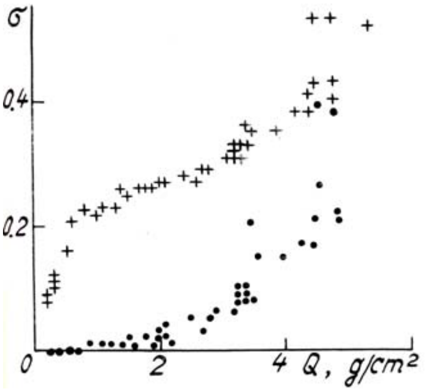

FIG. 2. The optimal values of the errors in determining the TOS for 48 regions from Ref. 9 for two values of σ_n: $\sigma_n = 0$ (dots) and $\sigma_n = 0.1$°K(crosses).

Figure 2 gives an Idea of the optimal values of σ for all 48 average atmospheric situations from Ref. 9 with $\theta = 60$'. The clear tendency for σ to increase as the integral moisture content of the atmosphere Q increases is interesting; for large values of Q (in the tropical regions), however, for the same values of Q σ can vary considerably from one region to another. This occurs because σ is determined not only by the integral content of water vapor in the atmosphere but also by the characteristics of the variability of its altitude distribution. The variability of the altitude profiles of the air temperature also has an effect.

The highest values $\sigma = 0.52$–0.53°K (with $\sigma_n = 0.1$°K) are reached in the Indian Ocean (region 4.3) during the spring, summer, and fall ($Q = 4.5$–5.4 g/cm^2). For those regions in which $Q < 3.5$–4.0 g/cm^2, $\sigma < 0.35$°K with $\sigma_n = 0.1$°K and $\sigma < 0.1$ K with $\sigma_n = 0$. This indicates that outside the tropics the quantity σ is determined primarily by errors in the measurements of the IR radiation.

Figures 3 and 4 show the optimal values of the coefficients α_1 and α_2 for each region. It should be noted that the optimal values of $\vec{\alpha}$ are different for different regions. The computational results obtained with $\sigma_n = 0$ are not presented in Fig. 4, since they fit completely into the region containing some of the points obtained with $\sigma_n = 0.1$ K, for which $\alpha_1 > 2.1$ and $\alpha_2 < -1.1$.

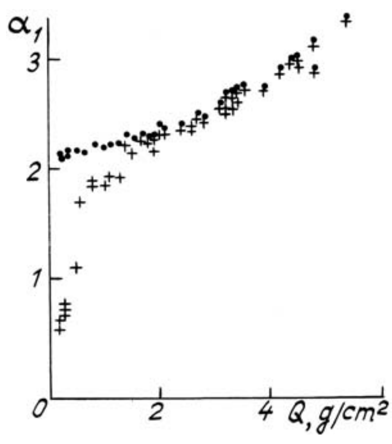

FIG. 3. The optimal values of the coefficient α_1 for 48 regions from Ref. 9 for two values of σ_n (the notation is the same as in Fig. 2).

Taking σ_n into account has virtually no effect on $\vec{\alpha}$ for large values of Q, but for $Q < 1$ g/cm^2 $\vec{\alpha}$ depends much more strongly on Q. It is obvious from the data shown in Fig. 4 (see Table I also) that for $\sigma_n = 0.1$°K for some regions (for which $Q < 0.3$ g/cm^2) even positive values of α_2 are obtained. This actually indicates that the two-angle method of atmospheric correction becomes degenerate; the optimal value of $\vec{\alpha}$ in such situations is determined not so much by the mechanism employed to take the atmospheric noise into account, as by the minimization of $\|\vec{\alpha}\|$, which determines the contribution of the errors in the measurements of T to σ. This effect is clearly illustrated by the data presented in Table I for region 2.6 (this region includes the Sea of Japan).

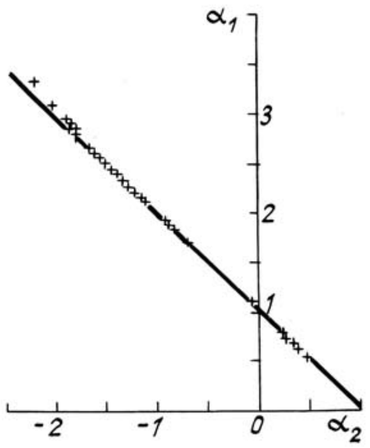

FIG. 4. The relation between the optimal coefficients α_1 and α_2 for regions from Ref. 9 with $\sigma_n = 0.1$ K. The strait line satisfies the condition $\alpha_1 + \alpha_2 = 1$.

We note that the estimates presented for the coefficients cannot be used in practice for a number reasons, the most important of which are the approximate character of the radiation model of SOA employed in the calculations (absolutely black surface, no clouds, no aerosol, etc.) and the a priori nature of the information.

The foregoing analysis demonstrates the approximate character of the simple linear method of atmospheric correction and makes it possible to understand more deeply the physical mechanisms responsible for the interfering action of the atmosphere in the determination of the TOS by remote methods from measurements at two angles. In different regions the errors in determining the TOS and the optimal coefficients in the algorithm for calculating the TOS differ significantly. The errors in the measurements of the IR radiation are an additional source of error. In most cases the TOS is determined with an accuracy of 0.2–0.35 K, but in separate situations it can be somewhat worse than 0.5 K.

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