RELATION BETWEEN THE DOWNWARD ATMOSPHERIC RADIATION IN THE INFRARED RANGE FROM 10 TO 12 μm AND THE INTEGRATED TRANSMITTANCE

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The relation between the integrated atmospheric transmittance P_0 and the downward and near–horizontal atmospheric radiation $I(0)$ and $I(90)$ is presented in the form of the expression $I(0)/I(90) = C_0 (1 - P_0)$, where $C_0 = 0.92 \dots 0.93$. It is based on the LOWTRAN–7 model calculations and on the analytical calculations using parametrizations for the vertical temperature profiles and atmospheric transmittance function. The absolute accuracy of determining P_0 according to this relation is estimated to be 0.02. The practical accuracy of determining P_0 from experimental values of $I(0)$ and $I(90)$ is equal to 0.035 ... 0.025 for $P_0 = 0.5 \dots 0.85$. The calculations of P_0 are described on the basis of the experimental data obtained in the water area of the Atlantic Ocean in spectral regions between 11 and 12 μm. The dependence of P_0 on the ambient humidity above the water surface is close to linear relation.

Of prime interest in thermal detection and ranging and remote temperature measurements is the problem of IR light propagation in so–called atmospheric windows where the transmittance is at its maximum and the distortion effect is correspondingly at a minimum. Experimental and theoretical studies of transmittance of the cloudless atmosphere in the IR region have been subjects of inquiry in Refs. 1–4. The transmittance is determined either from the measurements of solar radiation or with the use of multipass cells to give a measurement error of 0.02–0.05 (Ref. 2). Measurements of this kind involve sophisticated instrumentation which hinders experiments in a wide range of real meteorological conditions. At the same time extensive use is made of radiometers that are comparatively simple and highly sensitive to temperature variations.[5–7]

In this paper a procedure is proposed for evaluation of the atmospheric transmittance in the 800–1000 cm^{-1} range from the measurements of downward and near–horizontal radiation, $I(0)$ and $I(90)$.

MODEL ESTIMATES

Intensities of the downward zenith and near–horizontal radiation $I(0)$ and $I(90)$ in a spectral range of 800–1000 cm^{-1} (12.5–10 μm) with a step of 5 cm^{-1} were calculated using the LOWTRAN–7 code for three atmospheric models (tropics, midlatitude summer, and arctic summer). The relation between the integrated atmospheric transmittance P_0 and the ratio $I(0)/I(90)$ is depicted in Fig. 1. In practice, $I(90)$ can be replaced with $B(T_0)$, the Planck function of the temperature T_0 above the water surface, because $B(T_0) \approx I(90)$. The maximum deviations are found in the 980–1000 cm^{-1} spectral range, where the selective ozone absorption is most pronounced. This dependence on deviations of no more than 0.015 in the 800–980 cm^{-1} spectral range ($\lesssim$ 0.04 in the 980–1000 cm^{-1} range) is described by the relation

$$I(0) / B(T_0) \approx I(0) / I(90) = C_0 (1 - P_0), \quad (1)$$

where $C_0 = 0.93$.

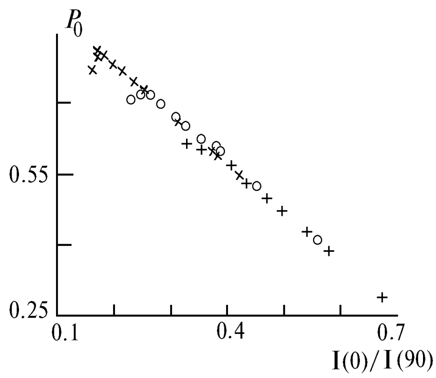

FIG. 1. The relation between the integrated transmittance P_0 and the intensity of the downward zenith radiation normalized to the intensity of near–horizontal radiation $I(0)/I(90)$ for three atmospheric models according to LOWTRAN–7: tropics (+ + +), midlatitude summer (∘ ∘ ∘), and arctic summer (× × ×) in the 800–1000 cm^{-1} spectral range with a step of 15 cm^{-1}.

Hence, the LOWTRAN–7 calculations show that, knowing the downward and near–horizontal radiation intensity (or the ambient temperature above the water surface) it is possible to determine the integrated atmospheric transmittance in the vertical column accurate to 0.02. Among other things, this furnishes the opportunity to assess the effect of different atmospheric constituents on the radiative transfer, using aerological measurements.

At the same time the real profiles of the transmittance function and temperature may disagree with the model presented in LOWTRAN–7. Let us consider a simple

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example of estimation of this disagreement taking into account the fact that a linear vertical temperature profile and exponential profiles of the absorbing constituents are characteristic of the atmosphere. Let the transmittance function P and the altitude temperature profile T be given by the relations:

$$P(z, 0) = P_0 + (1 - P_0) \exp(-\beta z) , \quad (2)$$

$$T(z) = T_0 + \alpha z . \quad (3)$$

where α and β are some constants, z is the altitude, and $P_0=P(\infty, 0)$ is the integrated atmospheric transmittance.

The radiative transfer equation normalized to $B(T_0)$ is of the form:

$$\Gamma = \frac{I(0)}{B(T_0)} = \int_0^\infty \frac{B(T(z))}{BT_0} \frac{\partial P(z, 0)}{\partial z} \, d z . \quad (4)$$

A distinguishing feature of Eq. (4) is that the integrand is of constant sign within the entire interval of integration and, consequently, the Γ value is monotone with respect to the parameters α and β. Thus, for any one of more complicated profiles of $T_c(z)$ and $P_c(z, 0)$ lying between the temperature and transmittance profiles given by $\alpha_1 < \alpha_2$ and $\beta_1 < \beta_2$ the relations $\Gamma(\alpha_1) < \Gamma_c < \Gamma(\alpha_2)$ and $\Gamma(\beta_1) < \Gamma_c < \Gamma(\beta_2)$ must be valid.

Substituting the value of $T(z)$ from Eq. (3) into the Planck function and considering the fact that all the absorbing constituents are found in the lower troposphere, i.e. within the range of integration $\alpha z < T_0$, we obtain

$$\frac{B(T(z))}{BT_0} - \frac{\exp\left(\frac{h c}{\lambda \kappa (T_0 - \alpha z)}\right) - 1}{\exp\left(\frac{h c}{\lambda \kappa T_0}\right) - 1} \approx \exp \frac{- h c \alpha z}{\lambda \kappa T_0^2}.$$

Passing to integration over P in Eq. (4) yields, in view of Eq. (2):

$$z = -\frac{1}{\beta} \ln \frac{P-P_0}{1-P_0} ,$$

and, correspondingly,

$$\Gamma = \int_{P_0}^{1} \exp\left[\frac{h c \alpha}{\lambda \kappa T_0^2 \beta} \ln \frac{P-P_0}{1-P_0}\right] d P(z, 0) = \frac{1 - P_0}{1 + \frac{h c \alpha}{\lambda \kappa T_0^2 \beta}}. \quad (5)$$

Eq. (5) may be written as

$$\Gamma = C (1 - P_0) , \quad (6)$$

where $C = C(\alpha, \beta, \lambda, T_0)$ and $C \approx 1$.

For representative values of $\alpha = 6$ K/km, $\beta = 1.1$ km^{-1}, $\lambda = 11.0$ μm, and $T_0 = 300$ K, the parameter $C = 0.92$, which agrees well with the previously calculated value of $C = 0.93$.

Eq. (5) agrees with exact solution (4) for $\alpha \to 0$ or $\beta \to 0$, i.e., for the isothermal atmosphere or provided that all the absorbing constituents are found near the surface. In these cases $\Gamma = 1 - P_0$, in virtue of Eqs. (4) and (5).

On a thorough inspection of Eq. (5), it should be noted that the parameter C has only a weak dependence on the wavelength. Thus, for the values of α, β, and T_0 given above $C = 0.923$ at $\lambda = 8$ μm and $C = 0.935$ at $\lambda = 12.5$ μm which for $C = 0.93$ yields $\delta P \le 0.07$, where δP is the error in estimation of P_0. Hence, Eq. (1) will be valid to given accuracy for any spectral interval within the 800–1000 cm^{-1} range under study.

It can be seen from Eq. (5) that for $\alpha = 0$–12 K/km, i.e. under conditions ranging from the isothermal atmosphere to that characterized by temperatures falling steeply with the altitude, for $P_0 = 0.5$–1 the maximum value of $\delta P \le 0.05$. Typical variations of $\beta = 0.4$–10 km^{-1} (the large values correspond to the cases where the main absorbing constituents are found near the surface, for example, in fog conditions) introduce errors of no more than 0.05 in the range cited. The computational error for P_0 using experimental data is estimated as

$$\frac{d P_0}{P_0} \approx \frac{1 - P_0}{P_0} \left(\frac{\delta C_0}{C_0} + \frac{\delta I(90)}{I(90)} + \frac{\delta I(0)}{I(0)}\right) + \frac{0.02}{P_0} .$$

The last term characterizes the accuracy of Eq. (1). The first term in parentheses is no more than 0.025. The performance parameters of the IR–radiometer enable us to determine $I(90)$ to within 1% and $I(0)$ to within 1.5, 4, and 6% for $P_0 = 0.5$, 0.7, and 0.85, respectively. Then $\delta P_0 \approx 0.035$, 0.03, and 0.025, and $\delta P_0/P_0 \approx 7$, 4, and 3% for the values of P_0 cited. Thus, the integrated transmittance of the atmosphere can be determined to sufficient accuracy from the measured values of $I(0)$ and $I(90)$ over a wide range of meteorological parameters.

EXPERIMENTAL INVESTIGATIONS

To perform two–channel measurements, a special experimental system was developed in the Radiometric Laboratory of the Marine Hydrophysical Institute of the Ukrainian Academy of Sciences. The setup includes control and recording devices. The viewing angle was varied between 0 and 180° from zenith by rotating the mirror. The mirror rotation allowed sighting of the calibration chamber. The radiometer design is similar to that described in Ref. 7.

The radiometer sensitivity range was determined by filters with centers at 11 and 12 μm and a transmission bandwidth of 0.3 μm.

The measurements were performed in the North Atlantic from 0 to 55° N in the North Sea, in the section from Great Britain to Boston through Atlantic, in the latitudinal section from Boston to 10° N, near the North–Western coast of Africa, and in the Mediterranean and Black Seas.

A specific example illustrating the dependence of the optical depth ratio $\tau(\lambda=11\ \mu m)/\tau(\lambda=12\ \mu m)$, $\tau(\lambda) = \ln(P_0(\lambda))$, calculated from the experimental data, on the humidity of the air above the water surface a (g/m^3) is given in Fig. 2. The constancy or weak variability of this ratio forms the basis of double–spectra procedures for retrieving the sea surface temperature.[8–10] Note that a mean ratio of 0.6 derived by Eq. (1) differs from a calculated value of 0.71 based on LOWTRAN–7 for standard atmospheres, which is due to the fact that the values of $P(0)$ calculated by the LOWTRAN–7 model for the 12.0 μm channel are overestimated as compared to those obtained experimentally under similar meteorological conditions.

Concurrent aerological observations and downward radiation measurements will make it possible to asses the quality of the LOWTRAN–type model calculations.

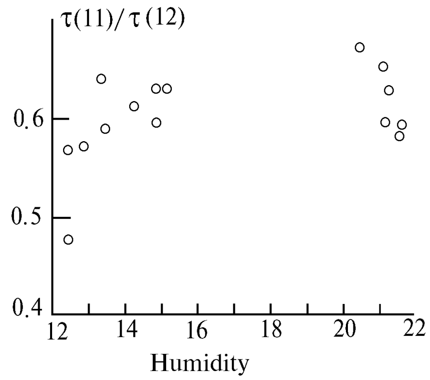

FIG. 2. The relation between the optical depth ratio $\tau(\lambda=11\ \mu m)/\tau(\lambda=12\ \mu m)$ *and the humidity of the air above the water surface a,* g/m^3 *as indicated by the* in situ *measurements.*

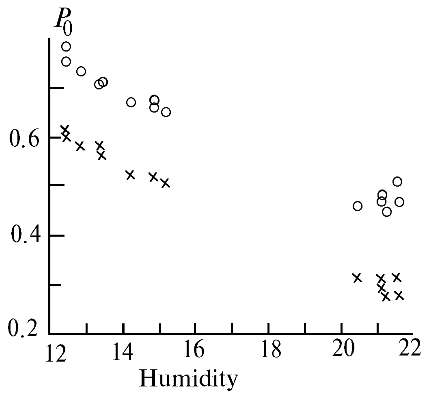

FIG. 3. The relation between the integrated transmittance P_0 *in the* 11–12 μm *spectral range and the humidity of the air above the water surface a,* g/m^3 *as indicated by the* in situ *measurements.*

Another applied aspect of Eq. (1) may be a search for the relationship between the transmittance P_0, the ambient temperature above the water surface and the humidity which would permit prediction of radiative transfer characteristics solely from meteorological measurements. Figure 3 depicts the dependence of the integrated transmittance P_0 in the 11–12 μm spectral range on the ambient humidity above the water surface, as indicated by the measurements made in the water area of the North and Equatorial Atlantic. It can be approximated as

$P_0=1.03 - 0.035a$ for the 12 μm channel,

$P_0=1.14 - 0.032a$ for the 11 μm channel.

This relationship between P_0 and a may be due to the self–similarity of the vertical profiles of a.[11]

CONCLUSION

We have proposed a procedure for determining the integrated atmospheric transmittance P_0 from measurements of the downward and near–horizontal radiation $I(0)$ and $I(90)$ using the LOWTRAN–7 – based numerical simulations and analytical calculations as well as parametrizations of vertical temperature profiles.

The P_0 calculations have been performed for the experimental data obtained in the water area of the Atlantic in the 11–12 μm spectral range. The value of P_0 appeared to be linearly related to the ambient humidity above the water surface for a ranging from 10 to 23 g/m^3.

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