SOME PROBLEMS IN DETERMINING OPTICAL THICKNESS OF THE ATMOSPHERE DUE TO EXTINCTION BY AEROSOL IN THE NEAR IR

D.M. Kabanov and S.M. Sakerin

Institute of Atmospheric Optics,
Siberian Branch of the Russian Academy of Sciences, Tomsk
Received February 24, 1997

The technique for determining the aerosol optical thickness of the atmosphere from the data of the transmission measurements in the wavelength range from 0.4 *to* 4 μm *has been analyzed. It is shown that the procedure usually used for taking into account (excluding) the gaseous absorption is incorrect and can lead to false results. The peculiarities are considered of taking into account the water vapor transmission function at its variable content, as well as the technique for calibrating the optical hygrometer on the basis of model spectroscopic data. We also discuss the errors in finding the aerosol component of the atmospheric transmission, and the results obtained in clear atmosphere over Atlantic ocean.*

1. INTRODUCTION

The method of spectral transmission based on photometric measurements of solar radiation propagated through the atmosphere is one of the most effective for investigation of the atmospheric aerosol and gas composition (m_2n, Cn_2, n_3, etc.). The major methodical problems in implementing this method are related to calibration and isolation of the components of the total transmission T_λ^0.

As known, the direct radiation attenuated by the atmosphere is determined by the total action of scattering and absorption effects

$$U_\lambda = U_{0\lambda}\, T_\lambda^0 = $$

$$= U_{0\lambda}\, T_\lambda^W\, T_\lambda^X\, T_\lambda^G \exp(-\tau^R m) \exp(-\tau^A m), \qquad (1)$$

where U_λ, and $U_{0\lambda}$ are the recorded and extraatmospheric values of solar radiation, T_λ^W, T_λ^X, and T_λ^G are the transmission functions of water vapor, ozone and other gas components, τ_λ^R, and τ_λ^A are the molecular (Rayleigh) and aerosol optical thicknesses (AOT), respectively, and m is the atmospheric thickness.

The AOT is usually found in the "transmission windows" of the atmosphere, which are relatively free of the effect of absorption by gases. But even in this case it is difficult to separate the aerosol component in the IR range due to the effect of gaseous absorption on the T_λ^0. The contribution of the majority of gases, T_λ^G, can be estimated from the model spectroscopic data for the average atmospheric conditions. Taking into account the absorption by ozone and, especially, by water vapor is not accurate enough even when models are used that take into account the season and regional peculiarities. Strong spatial and temporal variability of the atmospheric water vapor causes the necessity of simultaneous measurements of the total water content (TWC) of the atmosphere for the proper account of actual values T_λ^W.

One more difficulty is related to calibration (determination of signals $U_{0\lambda}$) and the procedure of calculating the absorption itself. The traditional method, the "long Bouguer method" (the "Langley plot method" (LPM) in foreign literature) is based on linear dependence of the logarithm of transmission on the atmospheric mass $\ln T_\lambda^0 = m\,\tau_\lambda^0$, i.e. it is assumed that $\ln T_\lambda^{W,G,X}$ also linearly depends on m, i.e. the absorption may be neglected. In this case one can find the value $U_{0\lambda}$ sought from linear extrapolation of $\ln U_\lambda(m)$ to $m = 0$. However, even if the spectral resolution is 1–5 cm^{-1}, one can hardly find a spectral interval completely free of absorption and satisfying the "linearity" condition. For example, according to data from Ref. 1, nonlinear dependence of $\ln U_\lambda$ on m can lead to a systematic error in determining $U_{0\lambda}$ up to 20%. The effect of "nonlinearity" is also noticeable at the next stage of excluding the gaseous component from the total optical thickness. It is clear that at a more coarse spectral resolution (using filters with $\Delta\lambda \cong 20$–200 cm^{-1}) one can not ignore absorption.

Taking into account the aforementioned facts, the techniCue for AOT measurements with a filter solar photometer[2] was developed. The idea of the method is in the application of the procedure of excluding absorption to the initial data, the signals U_λ. The application of the techniCue to shortwave range was considered earlier,[3] and this paper presents its further development. Let us note that other approaches to

0235-6880/97/08 540-06 $02.00

finding the aerosol component (see, for example, Refs. 4–7) are Cuite different, but they have one common peculiarity. The procedure for excluding absorption is applied in these methods only at the final stage after calibration and obtaining total effective optical thickness τ_λ^0. The uncertainty of such an approach is evident, but usually it is ignored because of other methodical problems in taking the absorption into account.

2. GENERAL STATEMENTS

The procedure for separating the aerosol component involves the following key elements:

1) The contribution of absorption is excluded by dividing the initial data U_λ by the values $T_\lambda^W T_\lambda^G$ and T_λ^X calculated for the experimental conditions;

2) the calibration procedure is applied to the corrected data, which depend linearly on m (for the convenience of subseCuent calculations the data are corrected taking into account the Rayleigh component T_λ^R too);

3) the values of the transmission functions T_λ^W, T_λ^X, and T_λ^G are calculated using the model[8] (program package) LOWTRAN-7;

4) the data from the spectral channel of the water vapor absorption at 0.94 μm are used when taking into account the variable value T_λ^W.

Preliminary analysis of the effect of the atmospheric conditions on the value of absorption showed that it is sufficient to perform continuously the account for the gas concentration variations, in the channels under consideration only for water vapor, i.e. one can assume the value T_λ^G (CO_2, CH_4, N_2O, etc.) to be constant within the limits of a region and season of the model[8]. In this case the values T_λ^R and T_λ^G depend only on the mass m and the instrumental function of the photometer.

$$T_\lambda^{RG}(m) = \left[\int_\lambda T_\lambda^{\text{ph}} \prod_i T_i^{RG}(m)\, \mathrm{d}\lambda\right] \Big/ \left(\int_\lambda T_\lambda^{\text{ph}}\, \mathrm{d}\lambda\right), \quad (2)$$

where T_λ^{ph} is the instrumental function of the photometer taking into account the spectral characteristics of the filter, receiver and source (Sun); T_i^{RG} are the transmission functions of the ith gas component set according to the model[8] taking into account T_λ^R.

For a convenience, the results of calculating T_λ^{RG} by EC(2) were approximated by the function

$$T_\lambda^{RG}(m) = \exp(-am^b). \quad (3)$$

The values of the approximation coefficients a and b for the midlatitude summer conditions are presented in Table I.

TABLE I. Parameters of the spectral channels and the approximations (3), (6), and (9).

Number of a spectral channel	λ_{max}, μm	$\Delta\lambda_{0,5}$, μm	a	b	c
1	0.369	0.020	0.4992	1	–
2	0.408	0.030	0.3211	1	–
3	0.423	0.012	0.2972	1	–
4	0.438	0.005	0.2534	1	–
5	0.484	0.007	0.1661	1	$1.693\cdot10^{-2}$
6	0.513	0.020	0.1303	1	$4.114\cdot10^{-2}$
7	0.558	0.016	0.0942	1	$9.118\cdot10^{-2}$
8	0.637	0.009	0.0583	1	$7.779\cdot10^{-2}$
9	0.671	0.010	0.0475	1	$4.156\cdot10^{-2}$
10*	0.940	0.010	0.0757	0.5096	–
11	0.871	0.019	0.0181	1	–
12	1.056	0.023	$7.46\cdot10^{-3}$	0.9905	–
13	2.182	0.029	$9.08\cdot10^{-3}$	0.7086	–
14	4.0	0.040	0.1398	0.8698	–

Below the peculiarities of taking into account the spatiotemporal variability of absorption by water vapor and ozone will be considered separately.

The data on T_λ^W, T_λ^X, and T_λ^{RG} make it possible to obtain the signals Y_λ free of the absorption effect

$$Y_\lambda = \frac{U_\lambda}{T_\lambda^W T_\lambda^{RG} T_\lambda^X} = U_{0\lambda} \exp[-\tau_\lambda^A m]; \quad (4)$$

$$\tau_\lambda^A = \ln[U_{0\lambda}/Y_\lambda]/m. \quad (5)$$

The value $\ln Y_\lambda$ linearly depends on m, so the correctness of the calibration of LPM is doubtless. One should note one positive factor more of the approach considered. Selection of the "calibration" days with the high and stable transmission T_λ^0 is important for determining $U_{0\lambda}$. It is obvious, that the diurnal variations of not only AOT but also of the water vapor (T_λ^W) automatically result in an increase of the error in determining U_0. Thus, by excluding T_λ^W at the first stage makes it possible to decrease not only the systematic error due to the "nonlinearity", but the random errors of calibration as well.

The technique of calibration using ordinary[9] or an iteration[7,10] method is not considered here.

3. ACCOUNT FOR THE ABSORPTION BY WATER VAPOR

When using solar photometers, the method of optical hygrometry[11, 12] is most effective for finding the total water content (TWC) and for taking into account T_λ^W. The peculiarities of its application are well studied, and the difficulties can be related only to the absence of balloon-borne data on humidity for calibration of the instrument. Investigations carried out in recent years[13–15] gave good grounds of the possibility of "theoretically" calibrating using model spectroscopic data. We have analyzed such an approach when using the spectroscopic data from the LOWTRAN-7 model. The calculations analogous to that presented in Ref. 14 show that the signal ratio in the range of water vapor absorption band $V = U_{0.94}/U_{0.87}$ can be presented in the form

$$V = V_0 T_W = V_0 \exp (a^* - b^* \sqrt{mW}), \qquad (6)$$

where $T_W = T_{0.94}^W / T_{0.87}^W$ is the ratio of the transmission functions calculated by formula analogous to EC.(2); $V_0 = (U_{0(0.94)}/U_{0(0.87)})\ (T_{0.94}^A / T_{0.87}^A)$ is the ratio of the "zero" signals taking into account small correction for aerosol (see Ref. 12 for a more detail), W is the TWC of the atmosphere, a^* and b^* are the parameters of the approximation, the values of which are presented in the row 10* of the Table I. Let us note that the function T_W has a more complicated form in the wide range $mW > 8$ than in EC.(6). In particular, the approximation was selected by the least squares method for finding the water vapor content W from the data $T_W = V/V_0$ (Fig. 1)

$$W = \left[d_0 + \sum_{i=1}^{3} d_i (\ln T_W)^i \right]^2 / m, \qquad (7)$$

where d_i are the coefficients calculated from the model for spectral channels (No. 10 and 11) of the photometer.

The unknown value V_0 can be calculated by the modified LPM[15] which is in extrapolation of the logarithm V as a function of $\sqrt{m}$ to $m = 0$. For calibration it is expedient to consider only the range $mW < 8$ where simple approximation (6) is fulfilled, and so the sought value V_0 is determined quite accurately.

For estimating the applicability of the technique developed, different comparisons were performed, including the final result, i.e. the data on water vapor content determined from the sounding balloon calibration[12] W_p and the model W_m. The results of measuring TWC over Tomsk in 1992 and 1995 were used for a comparison, and the relative error was estimated in the form $\delta_{p/m} = (W_p - W_m)/W_p$ (Fig. 2).

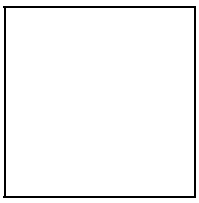

FIG. 1 Results of calculation of the transmission functions T_λ^W and approximations (7) and (8) for some spectral channels.

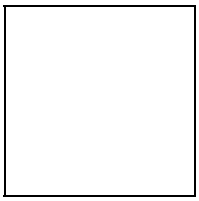

FIG. 2 Results of estimations of TWC of the atmosphere calculated for two techniques of calibration of the optical hygrometer.

As follows from the data presented the value of deviation does not exceed 5% in the measurement

range. The value $\delta_{p/m}$ increases up to 10% at $mW > 10 - 15$ (that corresponds to $W > 3$ g/cm^2, because $m < 4$). Analysis of the results shows that it is not the model, but the sounding balloon calibration that casts some doubt in the TWC range because of the absence of measurements at large values of mW. Thus, one can accept the value 5% as an upper limit of the model calibration error.

If the problem is to find T_λ^W and the not TWC, it is expedient to exclude the intermediate calculations and to find an expression for the transmission function directly in the form $T_\lambda^W = f(V, V_0, m)$. The approximation in the form of an exponential function was selected for compactness of this dependence that has the physical sense:

$$T_\lambda^W = \alpha + \sum_{i=1}^{2} \beta_i \exp\left[-\left(\frac{V}{V_0} - \eta\right) \Big/ \gamma_i\right]. \quad (8)$$

For brevity, let us present the values of the approximation coefficients in EC.(8) only for one spectral channel (2.18 μm): $\alpha = 1.027$; $\beta_1 = -0.2154$; $\beta_2 = -0.4254$; $\eta = 0.0886$; $\gamma_1 = 0.0590$; and $\gamma_2 = 0.3584$.

Since $V/V_0 \cong T_{0.94}^W$, the dependence (8) is representative of the relation between the water vapor transmission functions in different channels. The use of EC.(8) for the "aerosol" channels of the photometer makes it possible to control small variations of T_λ^W by the value T_W more sensitive to m_2n content.

The effect of the water vapor absorption in the visible channels (No. 1 to 7) is not significant, so it is taken into account by a simplified formula analogous to EC. (3) with the constant water vapor content (in the frameworks of the model), i.e. the functions T_λ^{GR} and the coefficients a and b for the channels 1 to 7 are calculated taking into account T_λ^W.

4. ACCOUNT FOR THE OZONE ABSORPTION

The procedure enabling accurate account for the ozone absorption T_λ^X in the channels of the visible range (the Chappui band) can be done analogously to the calculation of T_λ^W. However, due to the absence of the spectral channel adjusted to the ozone absorption band in our measurements and weaker effect of T_λ^X it was adopted to consider only the model values of the ozone content X. The variations of X were taken into account based on interpolation of the tabulated data of the two-dimensional model (latitude and month) presented in Ref. 16. Of course, the model selected corresponds only to the average conditions, but it is more thorough and concrete than the model from Ref. 8, where only two seasons and three climatic regions are considered.

Calculation of the function T_λ^X performed for different mW made it possible to obtain the expression analogous to EC. (3)

$$T_\lambda^X(m) = \exp\left[-c\,(mX)^{0.94}\right]. \quad (9)$$

The values of the coefficient c for the spectral channels 5 to 9 are presented in Table I.

5. ESTIMATION OF THE ERRORS

For the analysis of errors in τ_λ^A let us slightly change the form of the total transmission function of the gaseous components, and separate out the transmission function T_1 for a constant gas concentration and T_2 determining its variable part: T_λ^W T_λ^{RG} $T_\lambda^X = T_1\, T_2(W, X)$. Then, after differentiation of EC. (5) we obtain the expression for the errors $\Delta\tau_\lambda^A$ in the form

$$\Delta\tau_\lambda^A = \left[\left(\frac{\Delta U_0}{U_0\, m}\right)^2 + \left(\frac{\Delta T_1}{T_1\, m}\right)^2 + \left(\frac{\Delta T_2}{T_2\, m}\right)^2 + \right.$$

$$\left. + \left(\frac{\Delta U}{U\, m}\right)^2 + \left(\frac{\Delta m}{m}\,\tau_\lambda^A\right)^2\right]^{1/2} = \sum_{i=1}^{5} (\Delta_i)^2. \quad (10)$$

Two last terms (the errors in measuring U and determining m) are similar to that in the traditional techniCue[2, 9] being mainly random errors and can be estimated by the value of 0.0035 (at $m = 2$ and practically for any τ_λ^A). Three first terms characterize both the systematic and random errors. Let us analyze them from two point of view: A) estimation of the errors of the techniCue under consideration $\Delta_A^{(i)}$, and B) estimation of the differences in τ_λ^A in comparison with the data of calculations by the traditional techniCue $\Delta_B^{(i)}$. For brevity, let us present the results only for one spectral channel at 2.182 μm using the midlatitude summer conditions (Table II) as an example. Let us consider the components $\Delta_{A,B}^{(i=1-3)}$ separately.

TABLE II. Results of the estimation of the errors for the average conditions ($m = 2$, midlatitude summer).

Type of error	$i = 1$	$i = 2$		$i = 3$	$\Delta\tau_{2,18}^A$
		F	M		
$\Delta_A^{(i)}$	0.005	0.0016	0.0055 – 0.011	0.0019	0.009 – 0.013
$\Delta_B^{(i)}$	0.032		0.025	0.053	0.067

A) The LPM calibration error ($\Delta U_0/U_0$) is usually estimated[7, 9] to be about 1%, so $\Delta_A^{(1)}$ varies within the limits from 0.0025 to 0.01, and its contribution is maximum at $m = 1$.

One can isolate the error due to uncertainty in $T_\lambda^{ph} - \Delta_A^{(2ph)}$ and the error of the spectral model $\Delta_A^{(2M)}$ (see EC(2)) in the second term Δ_A^2. When estimating $\Delta_A^{(2F)}$ the displacement of the entire filter contour was simulated for the value of $\Delta\lambda = 16$ nm which is the maximum error of the spectral reference. The calculations showed that the effect of that kind of errors in the "transmission windows" is insignificant. The accurate calculation of $\Delta_A^{(2M)}$ in the particular spectral ranges is not a simple task and can be a subject for a separate study. The value $\Delta_A^{(2M)}$ is determined by the uncertainty of the spectroscopic data, the error in approximating the transmission function $T_i^G(m)$ and the difference of real atmospheric conditions (temperature, pressure) from the standard ones, for which the calculations were done. The effect of the latter factor dominates at the modern level of the spectroscopic calculations. For example, according to data from Ref. 17, one can estimate the error of the transmission functions due to the temperature profile variations to be the value of $\Delta(\ln T_\lambda)/(\ln T_\lambda) = \Delta T/T \ln T \cong 0.04$.

Although the reduced value characterizes the principal portion of the estimate, but it is not its total value. So, let us assume for better certainty that this error can lie in the limits from 0.04 to 0.08 in different spectral channels. Then the value $\Delta_A^{(2M)}$ under the average conditions, $m = 1$, is 0.008.

The last component $\Delta_A^{(3)}$ is determined (see EC(8)) by the errors in measuring the signals $V = U_{0.94}/U_{0.87}$ and calibration by the modified LPM. As before, it was assumed in calculations that $\Delta V_0/V_0 = 0.01$ and $\Delta V/V \cong 0.01$. Then one can obtain the value of $\Delta_A^{(3)}$ within the range 0.0014 to 0.0032.

Thus, the estimate of the total error $\Delta\tau_{2.182}^A$ is up to 0.013, and the systematic errors in calibration $\Delta_A^{(1)}$ and the spectral model $\Delta_A^{(2M)}$ make the principal contribution.

B) The difference between the results on main AOT, $\Delta_B^{(i)}$, is caused by the fact that the traditional techniCue does not take into account nonlinear dependence of $\ln U_\lambda$ on m (Fig. 3) due to the linear extrapolation of $\ln U_\lambda$ $m = 1$, 2–4 to $m = 0$. The error $\Delta U_{0\lambda}/U_{0\lambda}$ in the example shown in Fig. 3 is 0.065, and the corresponding value $\Delta_B^{(i)}$ lies in the limits from 0.016 to 0.065 for different m values.

In addition, the nonlinearity leads to incorrect account for the absorption at varying mass m. The calculations showed that maximum differences between the results obtained by two methods are observed at big m (for example, at $m = 4$ $\Delta_B^{(2)} \cong 0.04$).

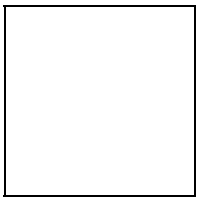

FIG. 3 Illustration of the calibration procedure for the signals $\ln U_\lambda$, $\ln Y_\lambda$ *and the resulting error in finding* $U_{0\lambda}$.

As to the last term $\Delta_B^{(3)}$, the errors were estimated for the case when real variability of TWC is not taken into account. It was assumed in the calculations performed for Western Siberia that the mean value of TWC in July is approximately 2.8 g/cm^2, and the rms error is 0.7 g/cm^2, and the full range of the variability is from 0.6 to 3.8 g/cm^2. The value $\Delta_B^{(3)}$ at such variations and $m = 2$ is 0.053, and the mean value is about 0.03.

The results of estimating the errors $\Delta_B^{(1-3)}$ confirm that the neglect of the nonlinearity and real TWC variations when determining AOT in the IR wavelength range can lead to a significant error (maximum up to 0.09). In addition, it should be noted that the effect of nonlinearity of the dependence $\ln U_\lambda = f(m)$ essentially weakens as the absorption in the visible wavelength range decreases. The estimates for this wavelength range showed that $\Delta_B^{(1-3)}$ does not usually exceed the value 0.01.

Let us present in conclusion an example of reconstruction of the data on τ_λ^{A1} obtained when studying the transmission of the atmosphere over the Central Atlantics in the fall of 1996 (Fig. 4). The results of calculations by the traditional procedure τ_λ^{A2} are also shown in this Figure for a comparison. The illustration using the marine atmosphere, as an example, with low aerosol content well characterizes the differences appearing due to incorrect account for the absorption.

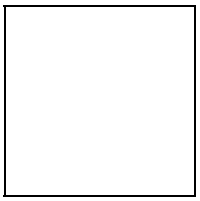

FIG. 4 Results on τ_λ^A obtained under the typical atmospheric conditions over the ocean.

As follows from the comparison made between τ_λ^{A1} and τ_λ^{A2}, the data coincide in the short wave part of the visible region. Small differences are observed in the range 0.52 to 0.7 μm (at the level of the error in determining AOT), which are caused by a more accurate account of (τ_λ^{A1}), spatial-temporal variability of the water vapor and ozone. The value τ_λ^{A2} in the IR range takes even negative values because of all the aforementioned factors were not taken into account, and the value of the discrepancy in the data increases up to 0.02–0.07, that is in agreement with the estimates of $\Delta_B^{(1-3)}$. The evident error in data on τ_λ^{A2} can be partially corrected in a usual procedure for calculating AOT by using real values of the water content. The effect of "nonlinearity" is not taken into account here, and, hence, AOT is systematically underestimated in the IR wavelength range.

REFERENCES

1. N.N. Paramonova, A.M. Brounshtein, and A.D. Frolov, Tr. Gl Geofiz. Obs., issue 406, 101–114 (1978).
2. *Regional Monitoring of Siberia. Pt. 2* (Publishing House of Institute of Atmospheric Optics, Tomsk, 1997), 295 pp.
3. A. Ignatov, I. Dergileva, Yu. Ratner, S. Sakerin, and D. Kabanov, in: *Proceedings of the IGARSS'94* (California, USA, 1994), pp. 1497–1499.
4. V.N. Aref'ev, Atm. Opt. **2**, No. 10, 878–894 (1989).
5. A.K. Shukurov, M.S. Malkevich, and A.I. Chavro, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **12**, No. 3, 264–271 (1976).
6. Yu.A. Pkhalagov and V.N. Uzhegov, Opt. Atm. **1**, No. 10, 3–11 (1988).
7. S.M. Sakerin, A.M. Ignatov and E.B. Shibanov, *Instrumentation and Techniques for Determining the Spectral Transparency from a Research Vessel "Ocean Space,B Atlantic-89 Experiment* (Moscow, dep. VINITI, August 7, 1990, No. 4496–1390), pp. 11–35.
8. F. Kneizys, G. Anderson, E. Shettle, et al., User's Guide to LOWTRAN-7, AFGL-TR-0177 (1988), 137 pp.
9. G.P. Gushchin, *Methods, Instruments and Results of Measurements of the Spectral Transparency of the Atmosphere* (Gidrometeoizdat, Leningrad, 1988), 200 pp.
10. A.M. Ignatov, I.L. Dergileva, S.M. Sakerin, and D.M. Kabanov, in: *Proceedings of the IGARSS'93*, vol. III (California, USA, 1993), pp. 1091–1093.
11. A.M. Brounshtein and N.V. Kazakova, Tr. Gl Geofiz. Obs., issue 237, 18–22 (1969).
12. D.M. Kabanov and S.M. Sakerin, Atmos. Oceanic Opt. **8**, No. 6, 442–446 (1995).
13. K.J. Thome, B.M. Herman and J.A. Reagan, J. Appl. Meteorol. **31**, No. 2, 157–165 (1992).
14. C.J. Bruegge, J.E. Conel, et al., J. Geophys. Res. **97**, No. D17, 18759–18768 (1992).
15. K.J. Thome, M.W. Smith, et. al., Appl. Optics. **33**, No. 24, 5811–5819 (1994).
16. D. Tanre, B.N. Holben, and Y.J. Kaufman, IEEE Trans. Geosci. Rem. Sens. **30**, No. 2, 231–248 (1992).
17. J.A. Reagan, K. Thome, and B. Herman, IEEE Trans. Geosci. Rem. Sens. **30**, 825–831 (1992).
18. D.M. Kabanov and S.M. Sakerin, Atmos. Oceanic Opt. **9**, No. 5, 415–420 (1996).