Pergamon

Adv. Space Res. Vol. 21, No. 3, pp. 439–442, 1998

Printed in Great Britain
0273-1177/98 $19.00 + 0.00

PII: S0273-1177(97)00926-5

SENSITIVITY STUDY OF THE ÅNGSTRÖM EXPONENT DERIVED FROM AVHRR OVER THE OCEANS

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ABSTRACT

Aerosol optical depth, τ^A, has been operationally retrieved from NOAA satellites over global oceans since 1989 using measurements in the Advanced Very High Resolution Radiometer (AVHRR) channel 1 ($\lambda_1=0.63\ \mu m$). Recently, the question has been raised on developing a two-channel procedure, to additionally retrieve the aerosol Ångström exponent, α, using measurements in AVHRR channel 2 (λ_2=0.83 μm). Here, we evaluate theoretically the expected uncertainty, $\delta\alpha$, resulting from atmospheric, oceanic, and radiometric errors. In general, $\delta\alpha$ is inversely proportional to τ^A, with the proportionality coefficient depending upon spectral channels used, sun-view geometry, and uncertainties in the model retrieval parameters. For the AVHRR, retrieval of α under typical oceanic conditions ($\tau^A<0.1$) is highly uncertain. Empirical analysis of AVHRR data is underway, and field programs are being conducted to quantify these errors observationally.

INTRODUCTION

Aerosol optical depth, τ^A, is customarily measured by sun photometers. Griggs (1975) proposed derivation of τ^A from satellite. This is a more complicated problem since upward radiance depends not only upon τ^A, but upon the aerosol single scattering albedo and phase function, and the surface bidirectional reflectance. All these factors are unknown and vary from one point to another. Radiances in AVHRR channel 1 have been used for τ^A retrieval over global oceans since 1989 (Rao et al., 1989; Stowe, 1991). The ocean reflectance is close to zero in this spectral range, and the aerosol phase function was a prescribed non-variable.

If aerosol size distribution obeys Junge' law, then spectral optical depth is approximated by Ångström's formula $\tau^A(\lambda)=\tau^A(\lambda_o)(\lambda/\lambda_o)^{-\alpha}$, where λ_o is a reference wavelength. The Ångström exponent can be estimated from measurements at two wavelengths as $\alpha=-ln(\tau^A_1/\tau^A_2)/ln(\lambda_1/\lambda_2)$. Deuze et al. (1988) and Kaufman et al. (1990) reported retrievals of α from AVHRR for very hazy cases ($\tau^A\approx 1$). Typically, aerosol content over the oceans is much lower ($\tau^A\approx 0.1$). Under such conditions, its contribution is more difficult to separate from the upward signal, especially in the presence of measurement errors and modeling uncertainties. The present study discusses the AVHRR potential for routine derivation of α over global oceans. The methodology used here has been developed and applied in the sensitivity study of chlorophyll retrieval from space (e.g. Gordon, 1981; André and Morel, 1989). Here we apply this methodology to estimate the accuracy in α.

SURVEY OF THE MEASUREMENTS OF τ^A AND α OVER THE OCEANS

Information on typical variability of τ^A and α is useful to constrain these parameters in the model sensitivity computations. Also, the range of natural variability of α gives an idea of the needed accuracy of its retrieval. A survey of sun-photometer measurements of $\tau^A(0.55\ \mu m)$ and α by Smirnov et al. (1995) shows that under clean maritime conditions, $\tau^A<0.1$. With a moderate intrusion of continental aerosols, $\tau^A\approx 0.2$-0.3. In extreme cases, e.g. Saharan dust outbrakes, τ^A may reach or even exceed 1.0. Over this range of τ^A, even over land, no data have been reported with $\alpha<-0.4$ or $\alpha>2.3$. Therefore, $\Delta\alpha=\alpha_{max}-\alpha_{min}\approx 2.7$. Separating α into at least 5 intervals therefore requires absolute accuracy in satellite retrievals of $\delta\alpha\sim\pm 0.25$.

MODEL OF THE SATELLITE SIGNALS AND TWO-CHANNEL ALGORITHM FOR α RETRIEVAL

Apparent reflectances measured by AVHRR are given by a linearized form of the single scattering approximation of the radiative transfer equation (Viollier et al., 1980; Gordon and Morel, 1983; Tanré et al., 1992)

$$\rho_1 \equiv \frac{\pi L_1}{F_{s1}\mu_s} = t_1^{oz}\times(\rho_1^R + \rho_1^A + T_1\rho_1^S)\ ; \qquad \rho_2 \equiv \frac{\pi L_2}{F_{s2}\mu_s} = t_2^{\Delta w}\times(\rho_2^R + \rho_2^A + T_2\rho_2^S) \qquad \textbf{(1)}$$

where ρ_i are reflectances defined by normalizing measured radiances L_i, $W\cdot m^{-2}\cdot\mu m^{-1}\cdot sr^{-1}$, to the effective solar constants of the channels ($i=1,2$), F_{si}, $W\cdot m^{-2}\cdot\mu m^{-1}$, and to actual illumination geometry ($\mu_s=cos\Theta_s$, $\mu_v=cos\Theta_v$; Θ_s and Θ_v are the solar and satellite zenith angles). ρ^S_i, ρ^R_i and ρ^A_i are the diffuse oceanic reflectances, and contributions to the signals from Rayleigh and aerosol scattering, respectively; t^{oz}_1 and $t^{\Delta w}_2$ are transmissions by ozone in channel 1, and water vapor in channel 2; T_i is total atmospheric path transmittance. The effect of the vertical distribution of water vapor with respect to the aerosol layer is described by a factor Δ: $\Delta=0$ (1) when water vapor is beneath (above) it, and $0<\Delta<1$ in intermediate cases. Under typical conditions, $\Delta\sim\frac{1}{2}$ (e.g. Kaufman et al., 1990; Tanré et al., 1992), but can be significantly different.

Rayleigh and aerosol contributions, ρ^R_i and ρ^A_i, are presented as (e.g. Viollier et al., 1980)

$$\rho_i^R = \frac{P^R(\chi)}{4\mu_s\mu_v}\times\tau_i^R \qquad \rho_i^A = \frac{\omega_i^A P_i^A(\chi)}{4\mu_s\mu_v}\times\tau_i^A \qquad \textbf{(2)}$$

where χ is the scattering angle, and P^R and P^A are the Rayleigh and aerosol phase functions:

$$P^R(\chi)=\frac{3}{4}\times(1+\cos^2\chi)\ ; \qquad P^A(\chi)=f\times\frac{1-g_1^2}{(1+g_1^2-2g_1\cos\chi)^{\frac{3}{2}}}+(1-f)\times\frac{1-g_2^2}{(1+g_2^2-2g_2\cos\chi)^{\frac{3}{2}}} \qquad \textbf{(3)}$$

and where ω^A is albedo of single scattering, and P^A is represented by a two-term Heney-Greenstein phase function (e.g. Sturm, 1981). For AVHRR channels 1 and 2, the dependence of the parameters f and g upon wavelength can be disregarded (e.g. Viollier et al., 1980). Oceanic diffuse reflectance, ρ^S, consists of two components: reflectances from oceanic underlight, ρ^S_U, and foam, ρ^S_F. Absorbing and scattering components of the transmittance are (Viollier et al., 1980; Tanré et al., 1992)

$$t_1^{oz}=\frac{1}{1+a_1^{oz}(mZ)^{b_1^{oz}}}\ ; \qquad T_1=\exp(-\frac{1}{2}m\tau_1^R)\ ; \qquad t_2^{\Delta w}=\frac{1}{1+a_2^{w}(mW\Delta)^{b_2^{w}}}\ ; \qquad T_2=\exp(-\frac{1}{2}m\tau_2^R) \qquad \textbf{(4)}$$

where $m=1/\mu_s+1/\mu_v$; Z ($cm\cdot atm$) and W ($g\cdot cm^{-2}$) are the column ozone and water vapor amounts; and (a^{oz}_1, b^{oz}_1) and (a^w_2, b^w_2) are channel-specific coefficients.

Aerosol contributions (path reflectances) are separated from Eq.(1) as:

$$\rho_1^A \equiv \frac{\omega_1^A P_1^A(\chi)}{4\mu_s\mu_v}\times\tau_1^A = \frac{\rho_1}{t_1^{oz}} - \rho_1^R - T_1\rho_1^S; \qquad \rho_2^A \equiv \frac{\omega_2^A P_2^A(\chi)}{4\mu_s\mu_v}\times\tau_2^A = \frac{\rho_2}{t_2^{\Delta w}} - \rho_2^R - T_2\rho_2^S \qquad \textbf{(5)}$$

Taking their ratio gives

$$\frac{\rho_1^A}{\rho_2^A} = \frac{\omega_1}{\omega_2}\times\frac{P_1^A}{P_2^A}\times\frac{\tau_1^A}{\tau_2^A} \approx \frac{\tau_1^A}{\tau_2^A} = \left(\frac{\lambda_1}{\lambda_2}\right)^{-\alpha} \equiv \epsilon; \qquad \alpha = -\Lambda\times\ln\left(\frac{\rho_1^A}{\rho_2^A}\right); \qquad \Lambda \equiv \left[\ln\left(\frac{\lambda_1}{\lambda_2}\right)\right]^{-1} \qquad \textbf{(6)}$$

where it has been assumed that differences between ω and P^A for the two AVHRR channels are negligible in the computation of α. Uncertainties in ρ_1, ρ_2, τ^R_1, τ^R_2, ρ^S_1, ρ^S_2, Z, W, and Δ result in errors in α. Assuming that errors introduced by these sources of uncertainty in Eq.(6) are small, and expanding appropriate terms into a Taylor series,

$$\delta\alpha=-\Lambda\times\left(\frac{\delta\rho_1^A}{\rho_1^A}-\frac{\delta\rho_2^A}{\rho_2^A}\right)\equiv-\frac{\Lambda}{\rho_1^A}\times(\delta\rho_1^A-\epsilon\delta\rho_2^A)\ ; \qquad \delta\rho_i^A\approx\frac{\partial\rho_i^A}{\partial\rho_i}\delta\rho_i+\frac{\partial\rho_i^A}{\partial\tau_i^R}\delta\tau_i^R+\frac{\partial\rho_i^A}{\partial\rho_i^S}\delta\rho_i^S+\frac{\partial\rho_i^A}{\partial W}\delta W+\frac{\partial\rho_i^A}{\partial\Delta}\delta\Delta+\frac{\partial\rho_i^A}{\partial Z}\delta Z \qquad \textbf{(7)}$$

Eq.(7) shows that the error in α increases: 1) as $\lambda_1/\lambda_2\rightarrow 1$; and 2) when aerosol reflectance, $\rho^A_1\rightarrow 0$ (proportional to $\tau^A_1\rightarrow 0$). In both cases, one has indeterminences of the type "$0/0$" ($ln(1)/ln(1)$ and $ln(0/0)$, respectively), so that the errors in α are amplified as these two limiting cases are approached. In other words, a two-channel scheme can not be used if it is close to degenerating into a single-channel one, or when amount of aerosol (optical depth) is small. In the next section, Eqs.(5)-(7) are used to derive the sensitivities to the uncertainties in different model and calibration parameters.

SENSITIVITY OF THE RETRIEVED α

Atmospheric and oceanic factors. Errors due to Rayleigh optical depth and oceanic diffuse reflectance:

$$\delta\alpha^R = \Lambda\times\frac{P^R(\chi)}{P^A(\chi)}\times\left(\tau_1^R - \epsilon\tau_2^R\right)\times\frac{\delta p_o}{p_o}\times\frac{1}{\tau_1^A}; \qquad \delta\alpha^S = \Lambda\times\frac{4\mu_s\mu_v}{P^A(\chi)}\times\left(T_1\delta\rho_1^S - \epsilon T_2\delta\rho_2^S\right)\times\frac{1}{\tau_1^A} \quad (8)$$

$\delta\alpha^R$ and $\delta\alpha^S$ are proportional to the relative uncertaintiy in the surface pressure, and to the absolute error in oceanic reflectance, respectively. The error in α gradually decreases as $\tau^A{}_1$ increases, so that when aerosol content becomes big enough, neither oceanic surface (which is not seen through a turbid atmosphere) nor Rayleigh scattering (which contributes to the satellite signal weakly) influence α retrievals.

Errors due to uncertainties in ozone and water vapor, $\delta Z/Z$, $\delta W/W$, $\delta\Delta/\Delta$:

$$\delta\alpha_{oz}=\Lambda a_1^{oz}b_1^{oz}(mZ)^{b_1^{oz}}\left(1+\frac{P^R(\chi)}{P^A(\chi)}\frac{\tau_1^R}{\tau_1^a}\right)\frac{\delta Z}{Z}; \qquad \delta\alpha_w=-\Lambda a_2^w b_2^w(mW\Delta)^{b_2^w}\left(1+\epsilon\frac{P^R(\chi)}{P^A(\chi)}\frac{\tau_2^R}{\tau_1^A}\right)\left(\frac{\delta W}{W}+\frac{\delta\Delta}{\Delta}\right) \quad (9)$$

Calibration. Relative uncertainty in gain, $\delta\gamma_i/\gamma_i$, translates to error in the retrieved α

$$\rho_i=\frac{\pi\gamma_i}{F_{si}\mu_s}\times(C_i-C_{io}); \qquad \delta\alpha_\gamma=\Lambda\times\left(\frac{1}{t_1^{oz}}\times\left(1+\frac{P^R(\chi)}{P^A(\chi)}\times\frac{\tau_1^R}{\tau_1^A}\right)\times\frac{\delta\gamma_1}{\gamma_1}-\frac{1}{t_2^{\Delta w}}\times\left(1+\epsilon\frac{P^R(\chi)}{P^A(\chi)}\times\frac{\tau_2^R}{\tau_1^A}\right)\times\frac{\delta\gamma_2}{\gamma_2}\right) \quad (10)$$

Similarity of Eqs.(9)-(10) stems from the identical influence of the uncertainties in absorbing gases, which are assumed to be above the scattering atmosphere, and calibration: both have a multiplicative effect on the satellite signal. Two asymptotic regimes are found:

1) when $P^A\tau^A{}_1 >> P^R\tau^R{}_1$, then $\delta\alpha$'s approach constants which are channel-geometry-error specific. In this case, the satellite signal almost fully originates from the aerosol, and a multiplicative error in upward radiance is proportionally translated to a multiplicative error in τ^A, since the model assumes a direct proportionality between τ^A and path radiance, and conseqently to an additive error in α, which is calculated from logarithms of τ^A; and

2) when $P^A\tau^A{}_1 << P^R\tau^R{}_1$, then $\delta\alpha$ are inversely proportional to aerosol loading. In this case, contributions from Rayleigh scattering and oceanic reflectance dominate, and multiplicative errors in satellite signal are similar to errors in the above two components -- cf. with Eq.(8).

NUMERICAL ESTIMATES FOR THE AVHRR

Estimates were done for $\Theta_s=60°$, $\Theta_v=0°$ ($\chi=120°$), assuming $\tau^R{}_1=0.06$, $\tau^R{}_2=0.02$; $a^{oz}{}_1=0.08$, $b^{oz}{}_1=1.07$, $a^w{}_2=0.07$, $b^w{}_2=0.5$, $Z=0.34$ $atm\cdot cm$, $W=3$ $g\cdot cm^{-2}$, $\Delta=0.5$ (Tanré et al., 1992). Underlight: $\rho^S{}_{U,1}\approx 15\cdot 10^{-4}$ for Case 1 waters typical of the open ocean (Ignatov et al., 1995); $\rho^S{}_{U,2}\approx 0$ (Ignatov, 1996); the foam reflectance is $\rho^S{}_{F,2}\approx\rho^S{}_{F,1}\approx 5\cdot 10^{-4}$ under wind speeds 6-7 $m\cdot s^{-1}$ (Koepke, 1984). Therefore, we use $\rho^S{}_1=2\cdot 10^{-3}$ and $\rho^S{}_2\approx 5\cdot 10^{-4}$.

Uncertainties: We introduce the following uncertainties in the parameters affecting $\delta\alpha$:
Rayleigh optical depth (surface pressure): $\delta p_o/p_o=0.01$ ($\delta p_o\approx 10$ $mbar$) (e.g. McClain et al., 1994).
Oceanic reflectance: $\delta\rho^{SF}{}_1=\delta\rho^{SF}{}_2\sim 5\cdot 10^{-4}$, coherent in the two channels, results from variable foam cover (unknown wind) - - if one ensures Case 1 waters in the retrieval point. Errors may be appreciably larger over Case 2 and intermediate waters. Variation of $\delta\rho^{SU}{}_1=10^{-3}$ ($\delta\rho^{SU}{}_2=0$) represent this source of uncertainty.
Ozone: $\delta Z=0.02$ $atm\cdot cm$ ($\delta Z/Z\sim 6\%$) (e.g. McClain et al., 1994).
Water vapor: $\delta W/W=0.2$ ($\delta W\sim 0.6$ $g\cdot cm^{-2}$ for $W=3$ $g\cdot cm^{-2}$), and $\delta\Delta/\Delta=0.5$ ($\delta\Delta\sim 0.25$ for $\Delta=0.5$).
Calibration: AVHRR is not calibrated onboard, and post-launch calibration provides typical accuracies of $\delta\gamma/\gamma\sim 5\%$ (Mitchell et al., 1995). We accept $\delta\gamma_1/\gamma_1=0.02$, $\delta\gamma_2/\gamma_2=0.02$, independent in the two channels.
All components of systematic error in α are listed in Table 1 for three values of τ^A. The next to last column shows arithmetic sum of the error components (the "worst case" scenario); the last column lists result of summing in a RMS sense.

Table 1. Components of systematic error in the Ångström exponent resulting from uncertainties in:

τ^A	$\delta\alpha_R$	$\delta\alpha_{SF}$	$\delta\alpha_{SU}$	$\delta\alpha_{oz}$	$\delta\alpha_w$	$\delta\alpha_\Delta$	$\delta\alpha_{\gamma1}$	$\delta\alpha_{\gamma2}$	$\Sigma\delta\alpha$	$\sigma\delta\alpha$
0.1	0.10	0.06	0.51	0.10	0.34	0.85	0.44	0.53	2.9	1.3
0.3	0.03	0.02	0.17	0.05	0.13	0.38	0.20	0.23	1.2	0.5
0.5	0.02	0.01	0.10	0.03	0.09	0.26	0.15	0.17	0.8	0.4

One can reduce uncertainty in α using a "physical calibration" based on "anchoring" satellite retrievals to ground-truth data in a few reference points. For the AVHRR, this procedure is helpful only for a restricted time, since concentration/vertical distribution of water vapor changes with respect to the aerosol, and the sensor's calibration drifts with time. The results from numerical weather forecast or additional measurements can provide information on e.g. surface pressure and wind speed, or column ozone/water vapor, which are, however, of secondary importance, and have their own errors. Significant contributions from the uncertainties in relative vertical distribution of water vapor with respect to aerosol, and from the oceanic reflectance in channel 1 are difficult to account for operationally.

CONCLUSION

Retrieval of the Ångström exponent under typical oceanic conditions ($\tau^A<0.1$) is highly uncertain from the present AVHRR. The accuracy becomes better as aerosol optical depth increases, yet probably allowing only separation of not more than two intervals in retrieved α for $\tau^A=0.5$. This preliminary conclusion, obtained under many serious assumptions, requires further checking with numerical simulations and with real satellite data, in conjunction with field experiments, e.g. TARFOX and ACE. It is felt, however, that further approaching reality hardly increases optimism, since other important factors, not considered in present study, such as direct glint (especially when the surface is roughned by wind), and multiple scattering, come into play. Random errors, resulting from the radiometric noise and digitization, must be considered, too. These can be suppressed by averaging and smoothing, however.

ACKNOWLEDGEMENT

This work was done when A.I. has been a University Corporation for Atmospheric Research visiting scientist at the SRL, on leave from the Marine Hydrophysics Institute, Sevastopol, Crimea, Ukraine. Support for A.I. and R.S. has come from NASA contract L-90987C, as part of L.S.'s co-investigator responsibilities on the EOS/CERES Science Team.

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