On the Interpretation of Atmospheric Turbidity Measurements

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ABSTRACT

Model calculations show that the aerosol particles within the lower troposphere usually contribute more than 80% to the total optical thickness of all particles within the atmosphere. For relative humidities higher than 99% within a thin layer of about 80 m thickness, the main contribution to the total optical thickness comes from this layer.

1. Introduction

The attenuation of visible solar radiation in the atmosphere is used as a measure of air pollution or sometimes as an indicator of lower stratospheric dust (e.g., Volz, 1970; Ellis and Pueschel, 1971; Dyer, 1974). However, the interpretation of attenuation measurements is problematic when the vertical distributions of the particles and the relative humidity are unknown. The present paper relates the optical thickness of all aerosol particles along a vertical path through the atmosphere to the vertical profiles of the particle number density and relative humidity. As a first approach to our problem the effects of multiple scattering and surface albedo on the optical thickness are not considered because we are not concerned here with cases involving estremely high turbidity.

In our model we assume three different aerosol types appearing in the following three main layers of the atmosphere:

Main layer 1) lower troposphere
Main layer 2) upper troposphere
Main layer 3) lower stratosphere.

TABLE 1. "Standard case" where no temperature inversion is found within the lower troposphere.

Particle number per unit volume	Relative humidity
Exponential decrease up to 5 km height (main layer 1)	Linear or exponential decrease up to 12 km
Linear decrease from 5 km to 12 km (main layer 2)	Constant value of 20% above 12 km
Constancy above 12 km (level of the tropopause) apart from a maximum around 18 km (main layer 3)	
No particles above 24 km	

For the lower troposphere (main layer 1) two distinct situations are also considered: 1) continental aerosol and 2) maritime aerosol. The models used for the vertical profiles of the particle number density and relative humidity will only approximate the main features of these profiles. Two main cases (Tables 1 and 2) have been considered; these are described briefly on this page.

2. Model equations

For all cases the optical thickness of all aerosol particles has been computed along a vertical path through the atmosphere using the relation

$$\tau_D = \ln\left(\frac{I_0}{I}\right) - \tau_R, \tag{1}$$

for normal incident radiation at a wavelength of 0.55 μm; τ_R is the optical thickness of the gaseous constituents of the atmospheric air for the whole atmosphere. Within the optical thicknesses τ_D and τ_R

TABLE 2. "Inversion case" with a temperature inversion in the troposphere.

Particle number per unit volume	Relative humidity
Constant value from the earth's surface to the inversion due to mixing of the air within this layer (main layer 1)	Linear or exponential increase from the earth's surface to the inversion
Rapid decrease due to large-scale subsidence immediately above the inversion, then a linear decrease up to 12 km (main layer 2)	Rapid decrease immediately above the inversion, then linear decrease up to 12 km height
Above 12 km same situation as in the standard case (main layer 3)	Constant value of 20% above 12 km

TABLE 3. Standardized extinction coefficient $\sigma_1(f_0,\lambda)$, reference relative humidity f_0, and fitted exponent $2\epsilon^*(\lambda)$ at 0.55 μm for the different aerosol types.

Aerosol type	σ_1 (cm²)	f_0 (%)	$2\epsilon^*$
Continental aerosol within the lower troposphere (main layer 1)[a]	2.99×10^{-10}	75	0.69
Maritime aerosol within the lower troposphere (main layer 1)[b]	2.51×10^{-9}	70	0.52
Aerosol within the upper troposphere (main layer 2)[c]	1.07×10^{-9}	80	0.56
Stratospheric aerosol (main layer 3)[d]	3.77×10^{-9}	20	—

[a] Power law size frequency distribution of the particles.
[b] Nearly log-normal size frequency distribution.
[c] Power law size frequency distribution for all particles with radii >0.1 μm in the dry state.
[d] Log-normal size frequency distribution.

irradiance losses due to both scattering and absorption are included. As an example ozone absorption is included in τ_R. I_0 is the incoming solar irradiance outside the atmosphere, and I the solar irradiance at the earth's surface. The optical thickness τ_D is defined as

$$\tau_D=\tau_{DI}+\tau_{DII}+\tau_{DIII}=\sum_{i=1}^{n}\tau_{Di}, \tag{2}$$

where τ_{DI}, τ_{DII} and τ_{DIII} are the optical thicknesses of the three main layers of the atmosphere, and τ_{Di} the optical thicknesses of the n sublayers into which the atmosphere has been divided for computation purposes.

In addition to the value of τ_D contributions to τ_D from one or a group of sublayers are of specific interest. In this context we will make use of the ratios

$$\tau_{Di}/\tau_D,\quad \sum_{i=k}^{l}\tau_{Di}/\tau_D,\quad \tau_{DI}/\tau_D,\quad \tau_{DII}/\tau_D$$

and τ_{DIII}/τ_D. These latter ratios are independent of the elevation of the sun above the horizon when a plane parallel atmosphere is a suitable approximation, and the influence of the atmospheric refraction can be neglected.

The formulas used for the calculation of the τ_{Di} will now be described. The geometrical thickness δ_i of the ith atmospheric sublayer and the height H_i of its top above the earth's surface are given by

$$\left.\begin{aligned}\delta_i&=H_0\ln\left[\frac{p_0-(i-1)\Delta p}{p_0-i\Delta p}\right]\\ H_i&=\sum_{j=1}^{i}\delta_j=H_0\ln\left(\frac{p_0}{p_0-i\Delta p}\right)\end{aligned}\right\}. \tag{3}$$

The constants $H_0=8\times10^5$ cm, $p_0=1000$ mb and $\Delta p=10$ mb are used. The mean extinction coefficient σ_{Di} of the aerosol particles within the ith sublayer is given by

$$\sigma_{Di}=\frac{\tau_{Di}}{\delta_i}=N_i\sigma_1(f_0,\lambda)\left(\frac{100-f_0}{100-f_i}\right)^{2\epsilon^*(\lambda)}. \tag{4}$$

Here N_i is the particle number per unit volume (cm³), $\sigma_1(f_0,\lambda)$ the standardized extinction cross section (cm²), f_i the relative humidity in the ith sublayer, f_0 the standard relative humidity, λ the wavelength of light, and $2\epsilon^*(\lambda)$ the so-called fitted exponent (Hänel, 1972, Table 4). σ_1 is a function of the aerosol

TABLE 4. Results for the standard case at a wavelength of 0.55 μm ($f_{12\text{ km}}=20\%$).

Case no.	f_{surface} (%)	$f_{5.1\text{ km}}$ (%)	$N_{5.1\text{ km}}$ (cm⁻³)	$N_{12.1\text{ km}}$ (cm⁻³)	$N_{17.7\text{ km}}$ (cm⁻³)	τ_D	100× $[\tau_{DI}/\tau_D]$* (%)	100× $[\tau_{DII}/\tau_D]$** (%)	100× $[\tau_{DIII}/\tau_D]$† (%)	100× $[(\Sigma\tau_{Di})_{1\text{km}}/\tau_D]$ (%)	Description of the model
						Maritime situation ($N_{\text{surface}}=250$ cm⁻³)					
1	70	49.2	11.6	0.02	0.083	0.0719	96.6	3.2	0.2	57.7	Linear humidity decrease in the troposphere [h_2(I) = 1 km]
2	70	49.2	11.6	0.1	0.413	0.0724	95.9	3.2	0.9	57.3	
3	70	49.2	11.6	0.5	2.067	0.0752	92.3	3.2	4.5	55.2	
4	50	37.5	11.6	0.02	0.083	0.0579	96.1	3.7	0.2	55.9	
5	90	60.9	11.6	0.02	0.083	0.1086	97.6	2.3	0.1	62.9	
						Continental situation ($N_{\text{surface}}=10\,000$ cm⁻³)					
6	70	49.2	77.4	0.02	0.083	0.267	94.2	5.7	0.1	63.4	Linear humidity decrease in the troposphere [h_2(I) = 1 km]
7	70	49.2	77.4	0.1	0.413	0.268	94.0	5.7	0.3	63.2	
8	70	49.2	77.4	0.5	2.067	0.271	93.1	5.6	1.3	62.6	
9	50	37.5	77.4	0.02	0.083	0.200	92.8	7.1	0.1	60.9	
10	90	60.9	77.4	0.02	0.083	0.472	96.5	3.5	0.0	69.7	
11	50	20.2	77.4	0.02	0.083	0.176	92.5	7.4	0.1	63.1	Exponential humidity decrease in the trophosphere [h_2(I) = h_5(I) = h_5(II) = 1 km]
12	70	20.3	77.4	0.02	0.083	0.204	93.6	6.3	0.1	67.2	
13	90	20.5	77.4	0.02	0.083	0.271	95.1	4.8	0.1	74.4	
14	50	37.5	10.5	0.02	0.083	0.101	98.0	1.9	0.1	85.3	Linear humidity decrease in the troposphere [h_2(I) = 0.5 km]
15	70	49.2	10.5	0.02	0.083	0.139	98.4	1.5	0.1	86.4	
16	90	60.9	10.5	0.02	0.083	0.267	99.1	0.8	0.1	89.2	

* Up to 5.1 km.
** From 5.1 to 12.1 km.
† Above 12.1 km.

TABLE 5. Results for the inversion case at a wavelength of 0.55 μm ($N_{12.1\text{ km}}=0.02$ cm^{-3}; $N_{17.7\text{ km}}=0.083$ cm^{-3}; $f_{5.1\text{ km}}=f_{12.1\text{ km}}=20\%$; the main layer I extends from the earth's surface to the top of the Kth sublayer just below the inversion; $H_6=0.435$ km, $H_{12}=1.023$ km).

Case no.	f_{surface} (%)	$f_{\text{inversion}}$ (%)	τ_D	100× [τ_{DI}/τ_D]* (%)	100× [τ_{DII}/τ_D]** (%)	100× [τ_{DIII}/τ_D]† (%)	H_k (km)	100× [τ_{DK}/τ_D] (%)	Description of the model
				Maritime situation ($N_{\text{surface}}=250$ cm^{-3}, $N_{5.1\text{ km}}=0.028$ cm^{-3})					
17	70	80	0.0718	98.9	0.9	0.2	12	9.7	Linear humidity increase below the inversion
18	70	90	0.0837	99.0	0.8	0.2	12	12.0	
19	70	95	0.0956	99.2	0.7	0.1	12	15.1	
20	70	99	0.1251	99.4	0.5	0.1	12	26.6	
21	70	99.5	0.1416	99.4	0.5	0.1	12	34.6	
22	70	99.9	0.2052	99.6	0.3	0.1	12	53.3	
23	70	80	0.0701	98.9	0.9	0.2	12	10.0	Exponential humidity increase below the inversion [y(I)=0.02]
24	70	90	0.0774	99.0	0.8	0.2	12	13.0	
25	70	95	0.0845	99.1	0.7	0.2	12	16.1	
26	70	99	0.1063	99.3	0.6	0.1	12	31.3	
27	70	99.5	0.1211	99.4	0.5	0.1	12	39.3	
28	70	99.9	0.1820	99.5	0.4	0.1	12	59.4	
				Continental situation ($N_{\text{surface}}=10\,000$ cm^{-3})					
29	70	80	0.335	92.3	7.6	0.0	12	9.4	Linear humidity increase below the inversion [$N_{5.1\text{ km}}=0.330$ cm^{-3}]
30	70	90	0.407	93.7	6.3	0.0	12	12.5	
31	70	95	0.487	94.7	5.3	0.0	12	16.8	
32	70	99	0.729	96.5	3.5	0.0	12	34.2	
33	70	99.5	0.897	97.1	2.9	0.0	12	44.7	
34	70	99.9	1.719	98.5	1.5	0.0	12	70.3	
35	50	60	0.230	88.8	11.1	0.1	12	8.5	Linear humidity increase below the inversion [$N_{5.1\text{ km}}=0.330$ cm^{-3}]
36	50	70	0.250	89.7	10.2	0.1	12	9.6	
37	50	80	0.280	90.8	9.1	0.1	12	11.2	
38	50	90	0.335	92.4	7.6	0.0	12	15.1	
39	50	95	0.396	93.5	6.5	0.0	12	20.7	
40	50	99	0.602	95.8	4.2	0.0	12	41.3	
41	50	99.5	0.761	96.6	3.4	0.0	12	52.6	
42	50	99.9	1.569	98.4	1.6	0.0	12	76.6	
43	70	80	0.186	80.4	19.5	0.1	6	15.9	Linear humidity increase below the inversion [$N_{5.1\text{ km}}=0.127$ cm^{-3}]
44	70	90	0.224	83.7	16.1	0.1	6	21.3	
45	70	95	0.269	86.4	13.5	0.1	6	28.5	
46	70	99	0.446	91.8	8.2	0.0	6	51.3	
47	70	99.5	0.593	93.9	6.1	0.0	6	63.5	
48	70	99.9	1.369	97.3	2.7	0.0	6	82.9	

* Up to H_k.
** From H_k to 12.1 km.
† Above 12.1 km.

size distribution and the mean complex refractive index of the aerosol particles at the relative humidity f_0. For visible radiation, the parenthetical expression with the power $2\epsilon^*(\lambda)$ almost equals the ratio of the total geometrical cross sections of all particles per unit volume of air at the relative humidities f_i and f_0. The variation of the total geometrical cross section with relative humidity, in fact, depends primarily on the chemical composition of the particles. Thus at $\lambda=0.55$ μm ϵ^* depends primarily on the chemical composition of the aerosol particles. If Rayleigh particles would predominate optically, ϵ^* would also depend on the size frequency distribution of these particles. The values of σ_1, f_0 and $2\epsilon^*$ at the 0.55 μm wavelength used for the calculations, are given in Table 3 for each aerosol type.

Within each of the atmospheric sublayers the particle number N_i per unit volume and the relative humidity f_i are constant. N_i and f_i are given by

$$\left.\begin{aligned} N_i&=N_0\{1+a(H_i-h_1)+b\exp(-H_i/h_2)\\ &\qquad+c\exp[-d(\ln H_i/h_3)^2]\}\\ f_i&=\tilde{f}_0+x(H_i-h_4)+y\exp(-H_i/h_5)\end{aligned}\right\}, \tag{5}$$

where N_0, $\tilde{f}_0$, a, b, c, d, x, y, h_1, h_2, h_3, h_4 and h_5 are constants for each of the three main layers of the atmosphere. Eq. (5) allows arbitrary linear and exponential changes of N_i and f_i with height H_i as well as the introduction of a maximum of N_i around the height $H_i = h_3$.

3. Results

Results are compiled in Tables 4 and 5 for both the maritime and the continental situations for the standard and inversion cases, respectively. The different models within these tables are characterized by relative humidities and particle numbers per cm^3 at the earth's surface and at specific heights. Also compiled are the total optical thickness τ_D of the aerosol particles within the atmosphere, the percentage fractions of τ_D coming from the different main layers, i.e., $100(\tau_{DI}/\tau_D)$, $100(\tau_{DII}/\tau_D)$ and $100(\tau_{DIII}/\tau_D)$, as well as the percentage fraction of τ_D coming from the lowest kilometer of the atmosphere, i.e., $100(\Sigma\tau_{Di})_{1\,\mathrm{km}}/\tau_D$, for the standard case. For the inversion case, the percentage fraction of τ_D coming from the single sublayer with the number K situated just below the inversion, i.e., $100(\tau_{DK}/\tau_D)$ is given instead of $100(\Sigma\tau_{Di})_{1\,\mathrm{km}}/\tau_D$. The following features are common to all cases:

1) The largest contribution to the total optical thickness τ_D comes from the lower troposphere (main layer 1).

2) The upper troposphere (main layer 2) usually does not contribute significantly to the total optical thickness τ_D.

3) The stratospheric aerosol does not contribute significantly to the total optical thickness τ_D except for those cases where the particle numbers around 18 km are larger than about 2 cm^{-3}.

The following additional features are of importance for the standard case (Table 4):

4) In general, the contribution of the first kilometer above the earth's surface to the total optical thickness τ_D is larger than 55%. This contribution increases with increasing relative humidity at the earth's surface.

5) The larger the height h_2 (I) of the homogeneous aerosol atmosphere, the smaller is the contribution of the lowest layers to τ_D (compare cases 6, 9 and 10 with cases 14, 15 and 16).

6) The differences between the cases with a linear humidity decrease within the troposphere (cases 6, 9 and 10) and those having an exponential humidity decrease (cases 11, 12 and 13) are small.

For the inversion case (Table 5) the most important additional features are:

7) The contributions of both the layer between the earth's surface and the inversion (main layer 1) and the sublayer just below the inversion, i.e., $100(\tau_{DI}/\tau_D)$ and $100(\tau_{DK}/\tau_D)$ increase strongly with relative humidity ($f_{\mathrm{inversion}}$) of the sublayer just below the temperature inversion.

8) The results for a linear humidity increase below the inversion (cases 17–22) and those for an exponential humidity increase (cases 23–28) do not differ very much.

9) The relative humidity f_{surface} at the earth's surface does not significantly influence the results (cases 29–34 and cases 37–42).

10) The height of the inversion above the earth's surface also does not significantly influence the results.

4. Conclusions

From the results, it would appear that the following features are important for practical applications:

1. Turbidity measurements from mean sea level allow conclusions only on the aerosol within the lowest troposphere.
2. The stratospheric aerosol usually cannot be determined by turbidity measurements from mean sea level.
3. When the relative humidity attains very large values, even within a thin layer, the main contribution to the total optical thickness of all the atmospheric aerosol particles comes from this layer.

REFERENCES

Dyer, A. J., 1974: The effect of volcanic eruptions on global turbidity and attempt to detect long-term trends due to man. *Quart. J. Roy. Meteor. Soc.*, **100**, 563–571.

Ellis, H. T., and R. F. Pueschel, 1971: Solar radiation: Absence of air pollution trends at Mauna Loa. *Science*, **172**, 845–846.

Hänel, G., 1972: Computation of the extinction of visible radiation by atmospheric aerosol particles as a function of the relative humidity, based upon measured properties. *Aerosol Sci.*, **3**, 377–386.

Volz, F. E., 1970: Atmospheric turbidity after the Agung eruption of 1963 and the size distribution of volcanic aerosol. *J. Geophys. Res.*, **75**, 5185–5193.