Preliminary application of lidar to measurements of the marine aerosol concentration in the atmosphere over the Gulf of Gdańsk at Sopot*

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Marine aerosol concentration
Size function
Lidar

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Abstract

This paper presents the results of preliminary measurements of aerosol concentrations above the Gulf of Gdańsk for various altitudes and for various distances from the shore. The measurements were carried out by means of the FLS-12 lidar system from the coast at Sopot in June 1992.

1. Introduction

In order to determine the evolution of such parameters as the concentration and distribution of aerosol sizes under various hydrometeorological conditions, it is necessary to perform time-consuming investigations over vast areas. Traditional methods of studying the marine aerosol *in situ* provide only general information about the magnitude of the particles. Furthermore, these methods do not enable definite conclusions to be reached about the influence of the microstructure of the near-water atmosphere layer and the dynamic conditions of the sea surface on marine aerosol concentrations

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or on the distribution of aerosol particle sizes. In order to deal with these complicated phenomena, active optical methods must be employed. These enable optical parameters of aerosol particles to be measured rapidly over a distance of many kilometers.

Lidar methods play a special role in marine aerosol investigations (Zuyev and Nøast, 1983). If the dynamic parameters of the atmosphere are known, the lidar, which works on various wavelengths, can be used to determine the distribution of aerosol sizes at various altitudes (Zuyev and Nøats, 1983; Klett, 1981). This method allows for quick, simultaneous measurement of aerosol concentrations and the determination of their size distribution functions under various hydrometeorological conditions.

On the basis of the above-mentioned parameters, inferences can be drawn about the subtle structure of the turbulent layer near the sea and about the microstructure of the meteorological field. Thus, the influence of these parameters on the dynamic processes of mass and energy exchange between the sea and the atmosphere can be determined. This paper presents the results of measurements of aerosol concentrations above the Gulf of Gdańsk carried out by means of the FLS-12 lidar system.

2. Apparatus

The FLS-12 lidar is a tunable laser system designed for the remote sensing of aquatic media and the atmosphere in the visible light spectrum.

A dye laser pumped by an excimer laser acts as the source of excitation radiation. The secondary emission of the sensed object is collected by an optical system that includes a telescope and polychromator, and is recorded by multichannel photoreceivers.

The FLS-12 is equipped with two receivers, A and B. Receiver A is a set-up for investigations into phytoplankton or crude oil fluorescence, while receiver B allows for the measurement of light intensity over a narrow spectral range defined by interference filters. Receiver B is applied only to atmospheric measurements and is equipped with eight filters allowing an arbitrary wavelength to be chosen from the visible light spectrum. The wavelength chosen depends on the type of dye used.

The operation of the FLS-12 and its data processing are controlled by an IBM-PC/AT computer. Special software simplifies the use of the FLS-12 in field measurements and in manual or automatic modes. A block diagram of the FLS-12 is shown in Figure 1.

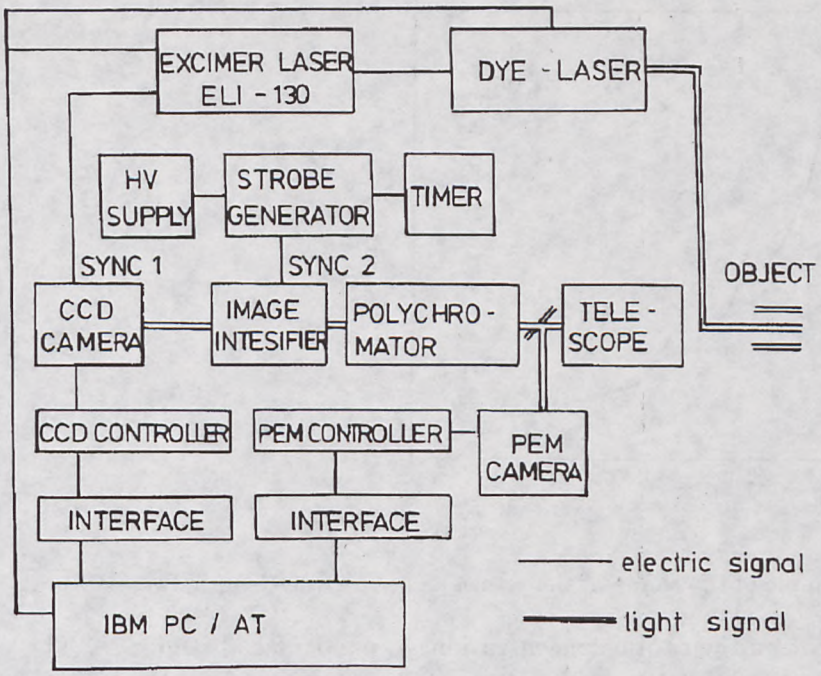

Fig. 1. Block diagram of FLS-12

3. Results and discussion

The field measurements were carried out on several nights in June 1992 in windless and cloudless weather. The atmosphere above the Gulf of Gdańsk between Sopot and Gdynia Orłowo was investigated. The air temperature T varied from 296 K to 298 K. Visible light of two wavelengths – $\lambda_1 = 450$ nm and $\lambda_2 = 562$ nm – was used for the measurements. Radiation was generated by a dye laser employing Coumarin 120 (λ_1) and Rhodamine 110 (λ_2) dye solutions. The power of the radiation produced, and hence the sensing distance, depended on the pulse generator voltage in the excimer laser. In the experiment, this voltage varied from 1.5 to 1.7 kV. The respective sensing distances were $z_{min} = 150$ m and $z_{max} = 420$ m, while the greatest altitude was $H_{max} = 62$ m. The distance $z_{max} = 420$ m corresponded to the atmospheric conditions as well as the lidar pulse power. The return signal from the atmospheric aerosol was recorded by the IBM-AT computer. An example of the shape of the return signal for wavelength $\lambda_1 = 450$ nm is shown in Figure 2.

For both wavelengths λ about a dozen return signals were registered. Those signals served as the basis for determining the distribution function

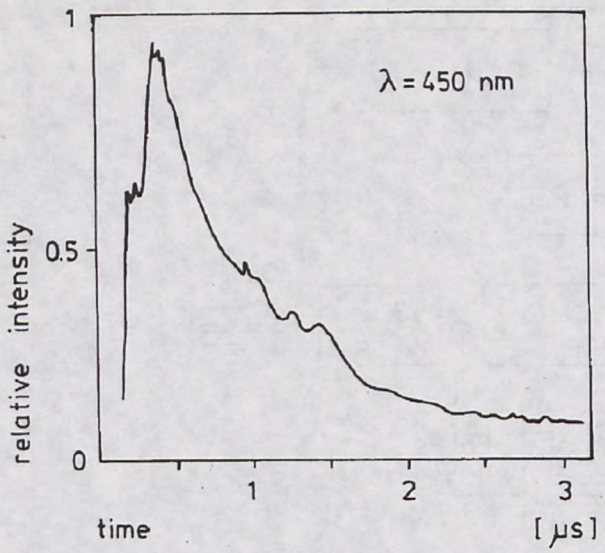

Fig. 2. An example of the shape of the return signal for wavelength $\lambda_1 = 450$ nm

of particle sizes and their total concentrations at particular altitudes H. The comparative method, introduced by Potter (1987), was used to derive these parameters. This method's principal hypotheses, listed below, coincide with the experimental conditions:

1. Numbers of wavelengths of laser radiation.
2. Extinction coefficient ratios are independent of the location of the target on the sensing beam route.
3. The extinction coefficient to back-scattering coefficient ratio is independent of the location of the target on the sensing beam route.

Using these guidelines, the lidar equation (*e.g.* Zuyev and Nøats, 1983) is as follows:

$$L_{ij}(R_i, \lambda_j) = K_j \alpha_{ij}(R_i, \lambda_j) \exp\left(-2 \int_0^{R_i} \alpha_{ij}(z_i, \lambda_j) dz_i\right), \tag{1}$$

where

$$L_{ij}(R_i, \lambda_j) = P(R_i, \lambda_j) \times R_i^2,$$

$$K_j = P_0(\lambda_j) \times T(\lambda_j) \times A \times B \frac{c(\tau + \tau_0)}{2},$$

i – describes the number of analysed points on the sensing beam route,

j – describes the number of wavelengths applied in the experiment,

$P_0(\lambda_j)$ – power of laser pulse for a particular wavelength λ_j,

$P(R_i, \lambda_j)$ – power of return signal of λ_j wavelength recorded from a distance R_i,
α_{ij} – atmospheric extinction coefficient at point R_i for wavelength λ_j,
A – effective area of telescope cross-section,
B – lidar efficiency,
c – speed of light,
τ – laser pulse duration,
τ_0 – counter opening duration,
$T(\lambda_j)$ – transmission of wavelength λ_j radiation through the atmosphere.

This lidar equation allows for the ready for determination of extinction coefficients in the following form:

$$\alpha_{ij}(R_i, \lambda_j) = \frac{x \times L_{ij}(R_i, \lambda_j)}{1 - 2x \int_{R_0}^{R} L_{ij}(z_i, \lambda_j) dz_i}, \tag{2}$$

where

$$x = \frac{1 - T_{ij}^2}{2 \int_{R_0}^{R_m} L_{ij}(z_i, \lambda_j) dz_i},$$

$$T_{ij} = \exp\left(- \int_{R_0}^{R_m} \lambda_{ij}(z_i, \lambda_j) dz_i\right),$$

R_0, R_m – distances from the lidar to the first and last point analysed.

The values of L_{ij} (see formula (2)) were determined at particular points with both wavelengths on the basis of the graphs generated by the return signals. Example values of atmospheric extinction coefficients $\alpha(r, \lambda)$ determined experimentally at two wavelengths, $\lambda_1 = 450$ nm and $\lambda_2 = 562$ nm, at various altitudes are shown in Fig. 3.

These values were used to determine the size distribution function of marine aerosol particles and their concentration at various heights. These parameters were determined using the procedure described by Piskozub (1991), where the extinction coefficient and the scattering properties of the aerosol are employed (see formula below):

$$\alpha_{ij}(R_i, \lambda_j) = \int_{r_1}^{r_2} Q(r_i, \lambda_j) \times S(r) dr, \tag{3}$$

where

$S(r)$ – total, geometric cross-section of aerosol particles per unit volume,
$Q(r, \lambda_j)$ – dimensionless extinction coefficient,
r – particle radius.

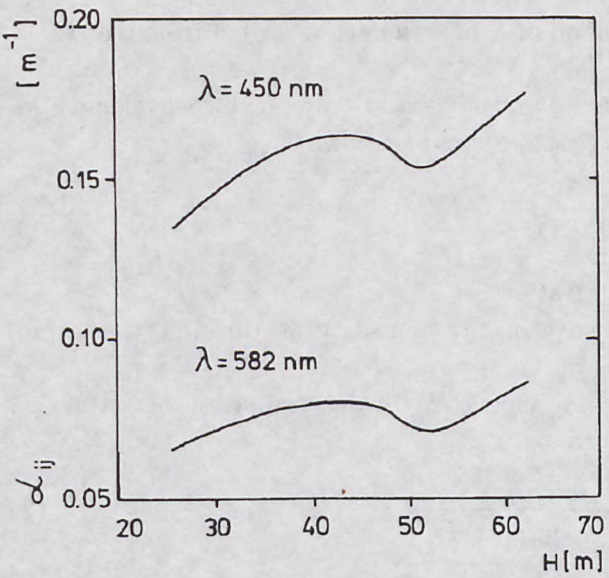

Fig. 3. Atmospheric extinction coefficient at various heights for two wavelengths

It was assumed that the marine aerosol is represented by non-absorbing spherical particles for which the density distribution function is exponential. Thus, functions $S(r)$ and $Q(r, \lambda_j)$ can be written as follows (Hulst, 1957):

$$S(r) = a \times r^2 e^{-br}. \tag{4}$$

$$Q(r, \lambda_j) = 2 - \frac{4}{\rho} sin\rho + \frac{4}{\rho^2}(1 - cos\rho), \tag{5}$$

where

ρ – $2 \times (m - 1)$,
m – relative light refraction coefficient,
x_j – Mie parameter, $x_j = \frac{2\Pi r}{\lambda_j}$,
a, b – distribution parameters.

The shapes of formulas 4 and 5 allow formula 3 to be solved in analytical form. Formula 3 can also be used to determine the distribution parameters a and b. These parameters are then matched with experimental values of coefficients $L_{ij}(R, \lambda_j)$. Calculated in such a way, coefficients a and b determine the function of aerosol size distribution and its concentration at particular altitudes

$$N_C(H) = \frac{a}{b} \qquad N_r(H) = N_C(H) \times f(r), \tag{6}$$

where

$f(r) = \exp(-br)$,
$f(r)dr = 1$.

The variations in mean aerosol concentration at various altitudes were determined on the basis of the many return signals recorded during the course of one hour (see Fig. 4).

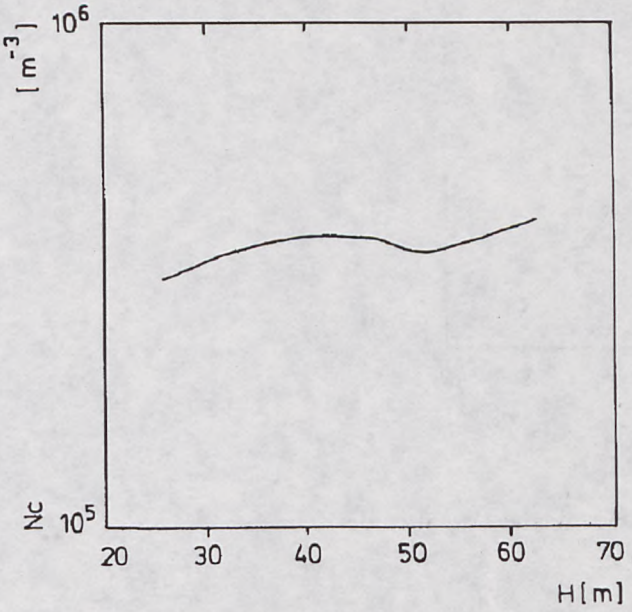

Fig. 4. The mean concentration of aerosol particles at various altitudes

Figure 5 shows concentrations obtained on the basis of the mean concentrations from Figure 4. Concentration variations are shown at two altitudes $H = 26$ m and $H = 62.5$ m for particle sizes in the range $(0.2 - 5\ \mu m)$.

Figure 5 shows that at each altitude, as the particle size increases, the concentration of aerosols decreases logarithmically. This is in agreement with the aerosol concentrations for windless weather obtained earlier (Garbalewski, 1980). The characteristic feature of this result is the inversion of the layers of higher aerosol concentration. It confirms the existence of stable stratification in the investigated air.

Figure 4, especially, reveals the existence of layers of various thicknesses and of various aerosol particle concentrations. On the basis of this figure one can distinguish four layers of different thickness and aerosol particle concentration: d_1 (40–26 m) 15 m; d_2 (50–40 m) 10 m; d_3 (58–50 m) 8 m and d_4 (70–58 m) 12 m. The results shown in Figure 6 confirm this division of the air.

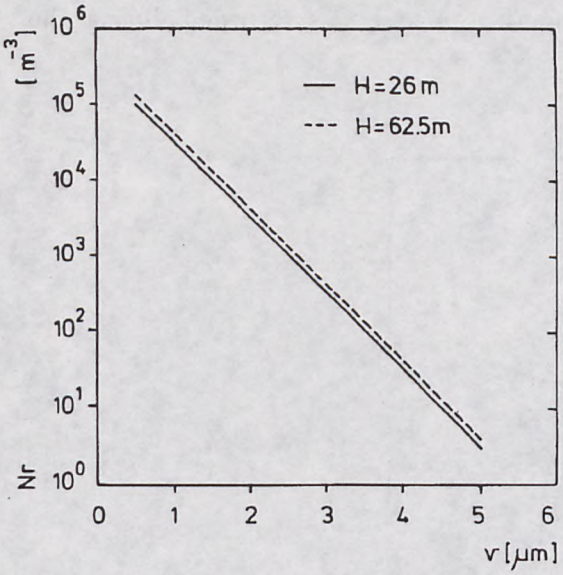

Fig. 5. Aerosol concentration *vs.* radius at two altitudes

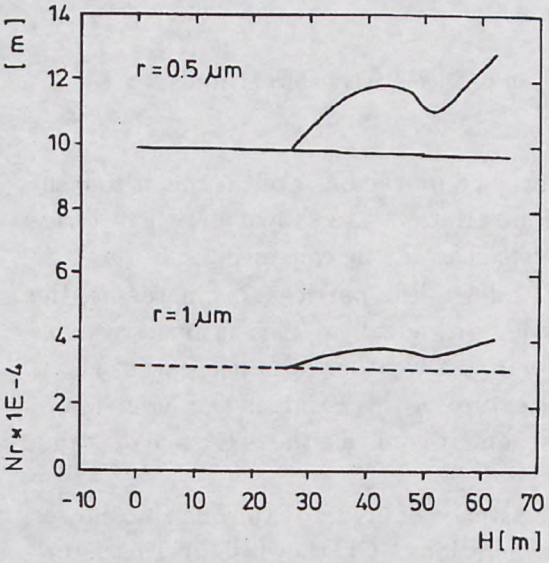

Fig. 6. Aerosol concentration of particles of selected size at various altitudes

Figure 6 shows the results of calculations, carried out for particle sizes $r_1 = 0.5\,\mu$m; $r_2 = 1\,\mu$m at various aerosol concentration levels and at various heights in a homogeneous atmosphere, obtained from the following formula:

$$N_T(r) = N_T^0(r) \exp(-\gamma \times H), \tag{7}$$

where

$N_T^0(r)$ – concentration of aerosol particles of radius r at height $H = 0$,

γ – coefficient whose value depends on r.

The values of the coefficient γ were taken from Cartney (1976). Comparison of the curves in Figure 6 reveals the existence of a layer-like structure of the air as well as the inversion of these layers. This is probably caused by an anthropogenic aerosol of size $r < 1\,\mu$m evenly distributed in these layers. Figure 7 shows changes in the total mass of marine aerosol per volume at various altitudes.

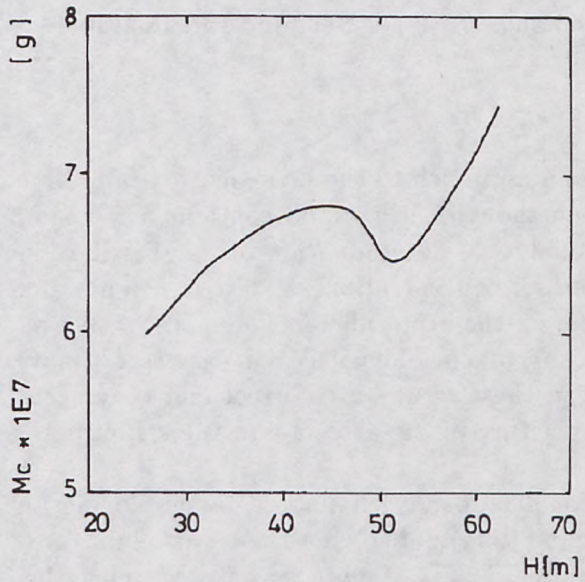

Fig. 7. Changes in masses of selected particles at various altitudes

Figure 8 shows changes in the masses of selected particles of $r = 0.5\,\mu$m, $r = 1\mu$m, $r = 3\,\mu$m at various altitudes.

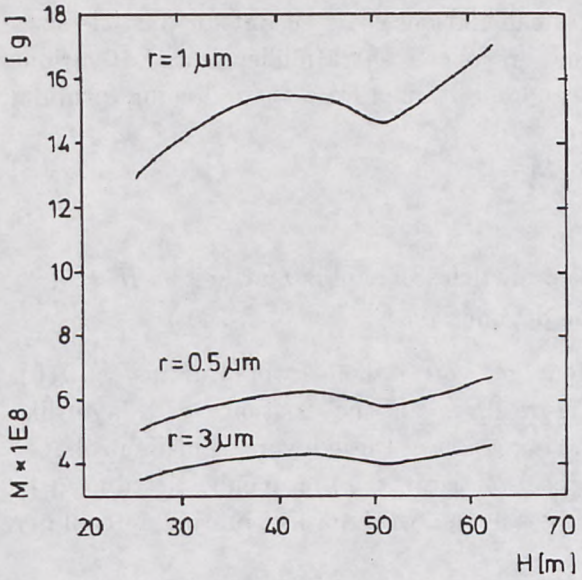

Fig. 8. Changes in total mass of marine aerosol per volume at various altitudes

4. Conclusions

The authors have presented an approach to the investigation of marine aerosols by means of lidar. Even though the method contains a few simplifications, it provides for the analytical determination of the distribution function of aerosol sizes and aerosol concentration at various heights and under various dynamic conditions of the atmosphere. The quantitative results obtained by the authors are erroneous and it is not easy to estimate this error. Qualitatively, however, these results are correct and they accurately describe both the microstructure of the aerosol and the structure of the atmosphere.

The source of these errors could be partly eliminated by employing an algorithm, introduced by Piskozub (1991), for solving the reverse lidar problem. In this method, which requires the use of more than five wavelengths, one can obtain the true spectrum of the optical properties of the medium being examined. Satisfactory results can be obtained even with only four wavelengths.

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