

VARIATION OF AEROSOL SIZE WITH ALTITUDE IN THE MARINE BOUNDARY LAYER INVERTED FROM MULTI-WAVELENGTH LIDAR BACKSCATTER DATA

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Abstract

In June, 1990, as part of the GLOBE-2 experiment, calibrated lidar backscatter was measured as a function of altitude at wavelengths of 0.532, 1.064, 1.54 and 9.25 μm . We have used inversions of the data collected 300 km west of Honolulu to obtain bimodal aerosol distribution models throughout the lowest 1280 m of the marine boundary layer. The models suggest that the coarse mode concentrations remain relatively constant with altitude, while changes in the fine mode concentration cause most of the variation in the total backscatter. The coarse mode, although contributing little to the total scattering area, has a significant effect on the lidar ratios and thus on the overall optical depth.

Introduction

Aerosols in the marine boundary layer (MBL) form an important component of the earth's climate system, as they result in radiative forcing which is opposite in sign to the forcing of anthropogenically added greenhouse gases such as CO_2 [1,2]. The vertical structure of the MBL is important in understanding the processes occurring within it, such as cloud formation, which have a large influence on radiative climate forcing [3]. Lidar is a well established tool for determining near-instantaneous vertical profiles of aerosol backscatter [4]. When such data are available at multiple wavelengths, in principle it becomes possible to invert for the corresponding aerosol size distributions.

In 1990, a unique experiment was performed using two pulsed lidar systems operating at wavelengths of 0.532, 1.064, 1.54 and 9.25 microns [5]. The systems were deployed on the NASA DC-8 during the Tokyo-to-Honolulu flight across the Western Pacific as part of the GLOBE-2 experiment [6, 7]. The data at 9.25 micron wavelength offered an opportunity to test previous conclusions [8] that such data would reduce the non-uniqueness of inverted aerosol size distributions.

Inversion Method

For the size distribution models, we assumed an M -modal lognormal distribution of spheres having radii r .

$$\frac{dN}{d \ln r} = \sum_{i=1}^M N_i \exp\left(\frac{-(\ln r - \ln R_i)^2}{2(\ln \mu_i)^2}\right) \quad (1)$$

m^{-3} , where N_i is the peak number concentration in m^{-3} , R_i is the central radius and μ_i is the spread in each mode. The backscatter at each wavelength was calculated by quadrature integration in $\ln r$ of Mie cross-sections for a sphere [9]. Details of the genetic inversion technique are similar to those used in previous papers [8, 10]. However, the fitness, F , which the genetic inversion attempts to maximize, is here defined by

$$F = \exp\left(-\sum_{k=1}^4 \left| \ln\left(\frac{d\sigma_{k,obs}}{d\Omega}\right) - \ln\left(\frac{d\sigma_{k,calc}}{d\Omega}\right) \right| / 4\right) \quad (2)$$

where $d\sigma_{i,obs}/d\Omega$ and $d\sigma_{i,calc}/d\Omega$ are the observed and calculated values, respectively, of differential lidar backscatter at the i th wavelength in $\text{m}^{-1}\text{sr}^{-1}$. $d\sigma_{i,calc}/d\Omega$ was obtained for multimodal lognormal size distributions of spherical particles by integrating their Mie cross-sections [8]. Refractive indices were estimated from values for hydrated sea-salt droplets [11, 12].

Data Description

Details of the two lidars used on the DC-8 flight are given in [4] and [5]. The lidar calibration factors were determined by pre-flight hard target calibrations [13]. Molecular absorption at 9.25 μm was estimated using AFGL and NECP aerosol models and techniques [14, 15]. The data were corrected for transmission losses using extinction to backscatter ratios calculated for these same models. A single set of data in a cloud-free region 300 km west of Honolulu (Figs. 1 and 2) was selected for this study.

To estimate the variation in hydration with altitude, needed for the refractive index calculation, we used relative humidities that were measured by a NOAA sonde at Lihue, 150 km east of the lidar sounding location (Fig. 3).

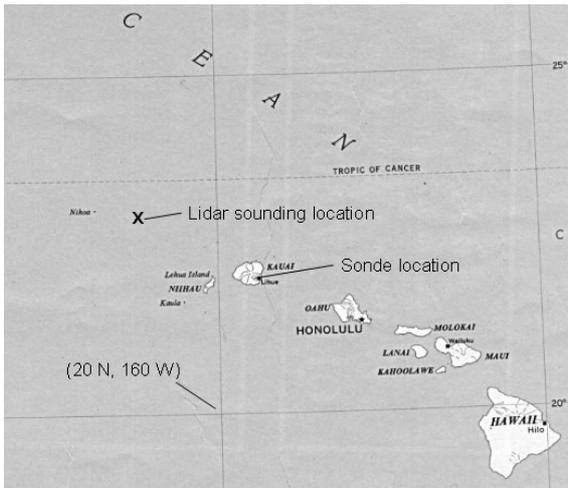


Figure 1. Map of the Hawaiian islands showing the location of the lidar sounding.

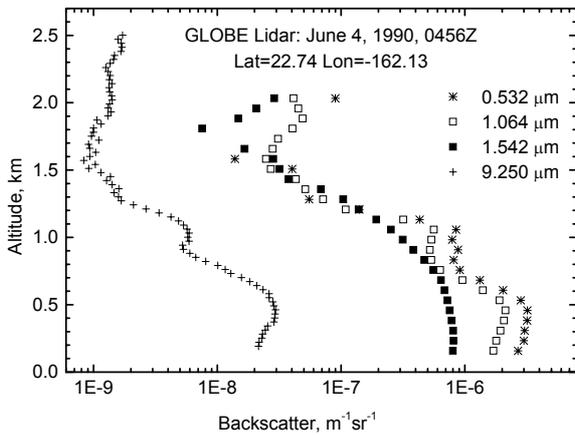


Figure 2. Variation of range-corrected backscatter with altitude.

Due to lower signal-to-noise on the 1.54 μm channel, the vertical averaging was higher, as is apparent from the data in Fig. 2. To allow for this, a weighting factor of 0.5 was applied to the 1.54 μm data in the subsequent inversions

Results

Fig. 4 shows the variation of measured backscatter (crosses) at the lowest altitude of 232 m. The theoretical curves were calculated for both bimodal and unimodal lognormal models (Fig. 5) obtained using genetic inversions of the data. Although the fits of both models to the data appears reasonable, the presence of the coarse mode has shifted the unimodal peak from 0.7 μm down to 0.2 μm . The large fluctuations in the theoretical backscatter between 2 and 17 μm are due to sea salt absorption lines [11].

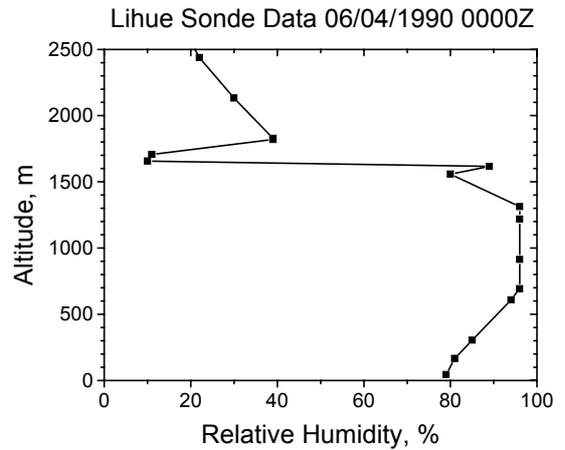


Figure 3. Variation of relative humidity with altitude measured at Lihue (Fig. 1) by a NOAA sonde released on 6/4/1990 at 0000 hrs UTC.

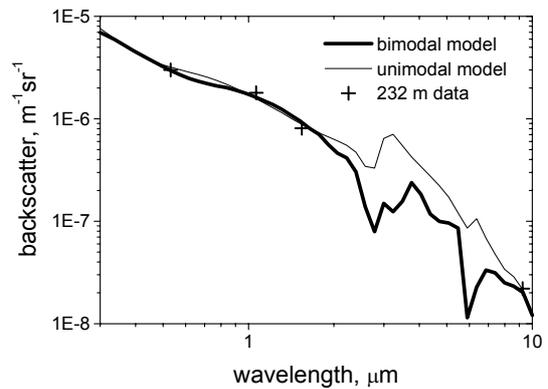


Figure 4. Measured GLOBE-2 backscatter (crosses) and variation predicted by the unimodal (dashed) and bimodal (solid) aerosol size distribution models in Fig. 5.

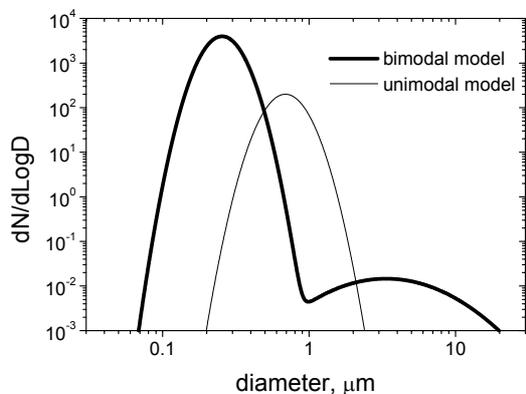


Figure 5. Unimodal (solid) and bimodal (dashed) size distributions inverted from the data in Fig. 4.

Inversions for bimodal models were performed on each of the four-wavelength backscatter data sets up to an altitude of 1280 m. Fig. 6 shows the resulting bimodal size distribution models plotted versus altitude, while Fig. 7 shows the errors of fit of the unimodal and bimodal models. The unimodal models do not fit the data as well as the bimodal models, particularly at the higher altitudes. Direct measurements of aerosol size distributions in the marine boundary layer [16] also indicate that in general, multimodal models are needed to realistically represent the contributions to scattering by marine aerosols.

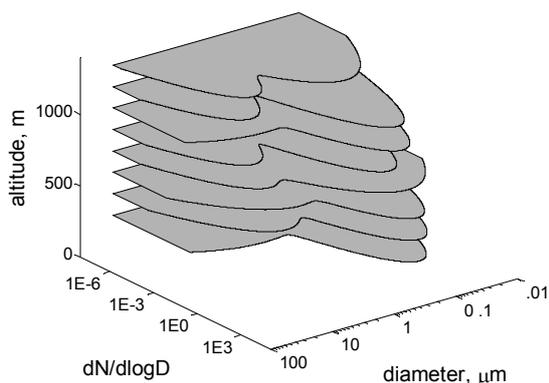


Figure 6. Bimodal aerosol distribution models inverted from the lidar data as a function of altitude.

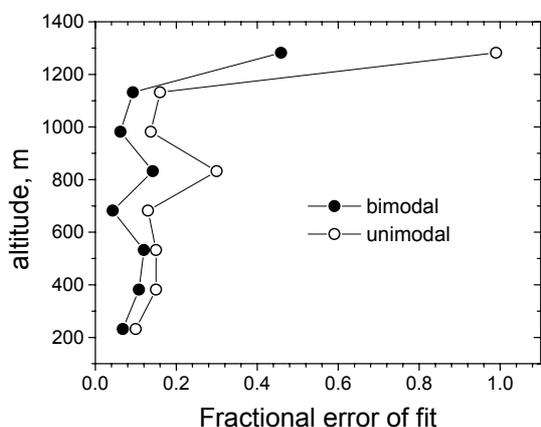


Figure 7. Comparison of fractional error of fit to data for unimodal and bimodal models versus altitude.

The genetic inversions were used to obtain multiple models fitting the data by using repeated runs with different random number seeds [8]. The spread of the resulting parameters then gives a quantitative estimate of the uniqueness of the models. The results for five repeat runs at each altitude are summarized in Figs. 8 and 9.

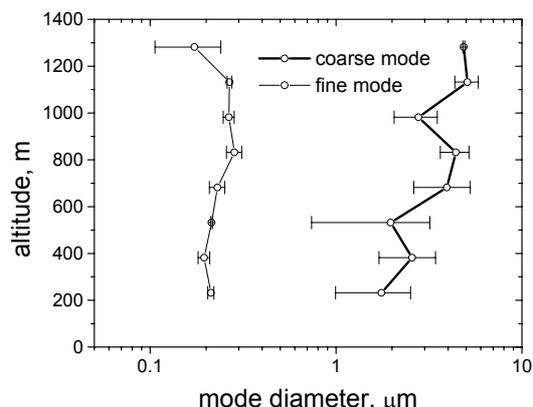


Figure 8. Mean and standard deviations of the mode diameters for five bimodal models at each altitude obtained using genetic inversions.

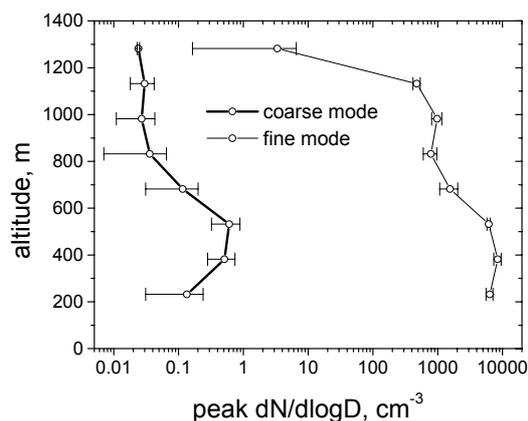


Figure 9. Mean and standard deviations of the peaks concentrations for five bimodal models at each altitude obtained using genetic inversions.

The results in Figs. 8 and 9 indicate a larger degree of ambiguity in the diameter and concentration of the coarse mode. Figure 10 shows the lidar ratios calculated for the models in Fig. 6.

Discussion

The variations of concentration with altitude suggest the presence of two regions, defined primarily by the fine mode concentration. The lowest region, which extends up to 532 m, has fine mode concentrations more than a factor of 10 higher than those measured at an altitude of 30 m at a coastal site north of Honolulu April, 2000 [17].

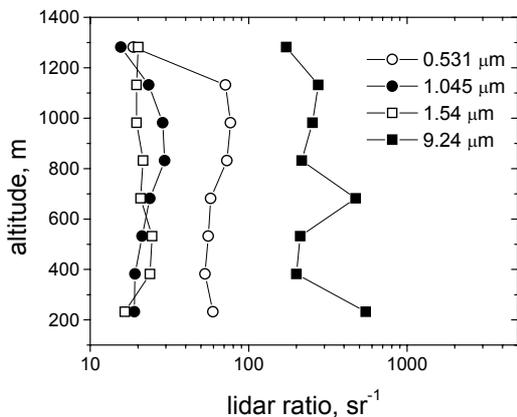


Figure 10. Variation of the lidar ratio with altitude predicted by Mie theory for the models in Fig 6.

However, the coarse mode concentrations at altitudes below 532 m (Fig. 9) are comparable to those measured during SEAS at 30 m, suggesting an additional source of fine-grained aerosols. The most likely source of these aerosols is volcanic ash from the Kilauea eruption [18], which commenced in 1983. The presence of volcanic ash is also consistent with the values of 50-60 for the lidar ratio at 0.53 μm (Fig. 10), compared to the values of 20-30 measured during SEAS [19].

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