Determinations of AOD in the Lower Troposphere and Stratosphere

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Abstract

Aerosol Optical Depth (AOD) is one of the most important parameters for explaining the distribution of aerosols within a column of air from the Earth's surface to the top of the atmosphere. Aerosols e.g., dust, smoke, pollution can block the solar radiation by either absorbing or scattering sunlight, causing direct and indirect effects to the Earth radiation balance. Two different methods are carried out in order to calculate AOD i.e., the analysis of obtainable data from Mie lidar measurement and the computational modeling with the mathematical assumptions. As a result, the determinations of AOD in lower troposphere and lower stratosphere are achieved for Sukhothai Province, Thailand ($17^\circ9’53”$N, $99^\circ51’43”$E). Analytical results of 5 consecutive days in January 2004 are considered into two cases corresponding to the atmospheric layers. In the lower troposphere, AODs are in the range of 0.475-1.240 and 1.698-1.719 with the RMSE and MBD of 1.118 and 1.556 from the analytical result of lidar data and the modeling, respectively. While, the lower stratospheric AODs are in the range of 1.657-1.712 with the RMSE and MBD
of 0.539 and 0.015. Hence, this work could successfully provide the columnar AOD determination by analytical and computational methods which are practical especially for the case of lacking of actual measurement physically and technically.

**Keywords:** AOD, Mie lidar, tropospheric aerosol, stratospheric aerosol

1. **BACKGROUND**

In general, aerosols are the small particles suspending in the atmosphere where they can exist in solid, liquid and semi-liquid forms (Boucher, 2015). These particulates are ubiquitous in air and often observable as dust, smoke, and haze. Both natural and human processes contribute to aerosols concentrations. On a global basis, aerosol mass derives predominantly from natural sources, mainly sea salt and dust. However, anthropogenic aerosols, arising primarily from a variety of combustion sources, can dominate the downwind of highly populated and industrialized regions, especially in the areas of intense agricultural burning. Aerosol loading, or amount in the atmosphere, is usually quantified by mass concentration or by an optical measure aerosol optical depth (Chin, 2009). Role of aerosols in climate and climate change is one of the largest uncertainties in our understanding of the present climate and in our abilities to predict future climate changes. Until the magnitude and the variability of anthropogenic aerosol radiative forcing are known, the predictions of future global warming may remain unacceptably high (Hobbs et al., 1997). The aerosols’ direct effect involves their interaction with solar and terrestrial radiation. The indirect effects include modification of the optical properties and life cycle of. Though the mechanism of aerosol direct effect is theoretically well understood, an assessment of its magnitude is more difficult due to spatial and temporal aerosol variability (Chylek et al., 2003). The aerosols interactions with the climate system in both troposphere and stratosphere are significant, particularly when the lifetime of stratospheric aerosols is rather longer than tropospheric ones (Boucher, 2015). In this paper we present the potential of computational modeling which can predict trend and uncertainty of size distribution and aerosol optical depths in an effort to determine aerosol optical properties when the actual measurement is unavailable and limited. Also, the accuracy of the AOD retrieval in two atmospheric layers was verified by two statistical tests.
2. METHODS

2.1 Study Area and Instrument
Sukhothai province (17° 9′ 53″ N, 99° 51′ 43″ E) locates in the lower northern part of Thailand, 440 km far-away from Bangkok. Its landscape is the basin of rolling plains with highlands and mountains, suitable for agriculture. So, it was selected as the study site because of its background and location in the Middle of Chao Phraya basin, representing the center of Thailand. Data of 5 consecutive days in January 2004 was analyzed in order to observe aerosol optical behavior during winter.

Mie lidar is energized by Nd: YAG laser with a wavelength of 532 nm. Linearly polarized laser pulses are vertically transmitted into the air, yielding the backscattering of signal from aerosols and clouds. Then, lidar measurement can provide the backscattered intensity, which contains the information of aerosols in the atmosphere.

2.2 Mathematical and Computational Algorithm

2.2.1 Analysis of Lidar Data
Basic theory of coaxial lidar (where the laser beam axis is parallel and close to the collecting mirror axis) governs with the relation of laser energy output \( P_0 \) and the backscattered signal \( P(z) \) (Biral, 2003), as

\[
P(z) = P_0 k \frac{c t A}{2 \ z} \beta(z) T(z)
\]

where \( k \) is a constant function of the intrinsic efficiency of the experimental apparatus, \( c t \) refers to the laser pulse length in atmosphere (the factor 2 refers to pulse round-trip), and \( A/z^2 \) is the solid angle comprised by the collecting mirror of area \( A \). The term \( \beta(z) \) is the volume backscattering coefficient, and the term \( T(z) \) refers to the transmissibility offered by the atmospheric path to photons traveling from the ground to a given distance \( z \). Normally, this attenuation term can be described as a negative exponential by the Bouguer – Lambert law (Rungjang, 2009) which is associated with the optical depth of the equation.

\[
T(z) = \exp(-\tau(z)) = \exp \int_{z}^{\infty} \alpha(z) \ dz
\]

where \( \tau(z) \) is denominated optical depth, and \( \alpha(z) \) is linear attenuation coefficient, since several terms in Eq (1) are constants, yielding the range corrected signal \( X(z) \).
as \[ X(z) = P(z) \ast z^2 = C \beta(z) \exp(-2 \int_0^z \alpha(z) \, dz) \] (3)

where \( C \) is the system calibration factor.

Additionally, from the light scattering theory lidar ratio of molecules \( (S_m) \), attenuation coefficients of molecules \( (\alpha_m) \) and backscattering coefficient of molecules \( (\beta_m) \) are related by \[ S_m = \frac{\alpha_m(z)}{\beta_m(z)} = \frac{8\pi}{3} \text{ sr}. \] The relationship between the attenuation coefficients of aerosols \( (\alpha_a) \), backscattering coefficient of aerosol \( (\beta_a) \). The lidar ratio of aerosol \( (S_a) \). The relationship is based on the equation (Biral, 2003)

\[ S_a = \frac{\alpha_a(z)}{\beta_a(z)} \] (4)

By the assumption of homogeneous atmosphere, the aerosol lidar ratio \( (S_a) \) of Sukhothai is 49.409 sr (Ruangrungrote , 2010).

From Fernald-Klett method, the volume backscattering coefficient of aerosols \( (\beta_a) \) is

\[ \beta_a(z) + \beta_m(z) = \frac{X(z)C(z)}{\beta_a(z) + \beta_m(z)} + 2S_a \int X(z)C(z)dz \] (5)

where \[ C(z) = \exp\left[ 2(S_a - S_m) \int \beta_m(z)dz \right]. \]

2.2.2 Aerosol Optical Depth (AOD)

The scattering and absorption of aerosol can be explained by the reduction of the incident light (attenuation) of aerosol particles (Ruangrungrote, 2012; Zielinski et al., 2014). The value of the attenuation coefficient over the entire vertical height \( z \) is called the aerosol optical depth (AOD), with the equation

\[ \tau_s = \int_0^z \alpha_s(z)dz \] (6)

where \( \tau_s \) is AOD and \( \alpha_s(z) \) is the attenuation coefficient along with the height between 0 and \( z \) (m\(^{-1}\)). So, AOD can indicate the amount of light lost by an attenuation of the light beam when light passes through aerosol particles.
2.2.3 Bird Model

$T_a(z)$ relates to AOD as seen in the equation (2) and equation (6), then Bird model is applied for determining $T_a(z)$ (Behar et al., 2015) as

$$T_a(z) = \exp\left[-L_{ao}^{\beta_i}(1+L_{ao} - L_{ao}^{\beta_j})m_{ao}^{1.253}\right]$$

(7)

where $L_{ao}$ is the broadband AOD with the following equation:

$$L_{ao} = \beta_i\left[0.695 + (0.016 + 0.066\beta_i (0.7)^{\beta_j})m_{ao}\right]^{\beta_j}$$

(8)

where $\beta_i$ is angstrom exponent, $\beta_j$ is Angstrom turbidity coefficient and $m_{ao}$ is air mass at actual pressure with the equation $m_{ao} = m_i \frac{P}{P'}$ where $P$ is actual pressure (mbar), $P'$ is standard pressure (mbar). And $m_i$ is air mass at standard pressure, written as

$$m_i = \left[\cos \theta_z + 0.15(93.885 - 0_z)^{-1.253}\right]^{-1}$$

(9)

However, this model needs some astronomical data as follows:

Solar zenith angle ($\theta_z$) is the angle between the observer and the sun, calculated by

$$\cos \theta_z = \sin \delta \sin \phi + \cos \delta \cos \phi \cos \omega,$$

where $\delta$ is the declination angle (degree), $\phi$ is the latitude of the location (degree) and $\omega$ is hour angle (degree).

Declination angle ($\delta$) is affected by the tilt of Earth's axis of rotation. The angular position of the sun at noon compared to the plane of the equator. The change is -23.45 to +23.45 degree, and in the form $\delta = 23.45\sin\left(\frac{360 \cdot 284 + n_j}{365}\right)$, where $n_j$ is number of the day in a year, $n_j = 1$ for January 1st.

Hour angle ($\omega$) is the hours of sun position in the sky at different time, written as $\omega = 15 \times (ST - 12)$.

Solar time (ST) is related to the difference between the longitude and longitude of
the observer, as the form $ST = SDT + 4(L_{st} - L_{loc}) + E$, where $SDT$ is local time for Thailand (hr), $L_{st}$ is longitude local standards for Thailand, equal to 105° E, $L_{loc}$ is local longitude of the position (degrees), and $E$ is the equation of time, $E = 9.87\sin 5B - 7.5\cos B - 1.5\sin B$, where $B = \frac{360(n - 81)}{365}$.

2.3 Statistical Validation

2.3.1 Root Mean Square Error (RMSE)
RMSE has been used as a standard statistical metric to measure model performance in meteorology, air quality, and climate research studies (Chai & Draxler, 2014). RMSE is less that the model used to estimate the study was close to the real. And if it is equal to zero, it means no errors in the model. The number of sample is $n$ samples.

$$
RMSE = \sqrt{\frac{1}{n} \sum_{i=1}^{n} (T_i - \tau_i)^2}
$$

where $T_i$ is AOD determined by the modeling, and $\tau_i$ is AOD from the analysis of lidar data.

2.3.1 Mean Bias Difference (MBD)

Most accurate model should have an MBD value close to zero (Behar et al., 2015).

$$
MBD = \frac{1}{nm} \sum_{i=1}^{n} (c_i - m_i)
$$

where $m_i$ and $c_i$ stand for the measured and predicted values of AOD, respectively.

Mean values of the measured and predicted values are defined as $m = \frac{1}{n} \sum_{i=1}^{n} m_i$.

3. RESULTS

Figure 1-3 present the vertical profiles of range corrected signal, backscattering coefficient and extinction coefficient of 5 consecutive days during 13-17 January 2004. The results of 24-hr AOD retrieval from modeling are shown in Figure 4. Then the period of daytime from 8.00 – 16.00 LCT was considered in order to narrow down the lower tropospheric and stratospheric AOD tendency as seen in Figure 5. Moreover, Figure 6 and Table 1 provide the comparison of AODs retrieved by the analysis of lidar data and computational model, together with the statistical tests.
Figure 1. The vertical profile of range corrected signal, backscattering coefficient and extinction coefficient of 13-14 January 2004 at 12.00 LCT.
**Figure 2.** The vertical profile of range corrected signal, backscattering coefficient and extinction coefficient of 15 - 16 January 2004 at 12.00 LCT.

**Figure 3.** The vertical profile of range corrected signal, backscattering coefficient and extinction coefficient of 17 January 2004 at 12.00 LCT.
Figure 4. Determinations of AOD from modeling.
Figure 5. Daytime AOD from 8.00 – 16.00 LCT in (5a) the lower tropospheric and (5b) lower stratospheric, obtained by the analysis of lidar measurement.

Figure 6. The comparison of AOD in the lower tropospheric and lower stratospheric, obtained by and the analysis of lidar measurement and the computational model.
Table 1. Average AOD in the lower tropospheric and lower stratospheric with the statistical tests

<table>
<thead>
<tr>
<th>Date</th>
<th>LT AOD</th>
<th></th>
<th>Statistical test</th>
<th>LS AOD</th>
<th></th>
<th>Statistical test</th>
</tr>
</thead>
<tbody>
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<td></td>
<td>lidar</td>
<td>model</td>
<td>RMSE</td>
<td>MBD</td>
<td>lidar</td>
<td>model</td>
</tr>
<tr>
<td>13/1/2004</td>
<td>0.572</td>
<td>1.719</td>
<td>1.258</td>
<td>2.004</td>
<td>1.668</td>
<td>1.719</td>
</tr>
<tr>
<td>14/1/2004</td>
<td>0.875</td>
<td>1.714</td>
<td>1.054</td>
<td>0.958</td>
<td>1.712</td>
<td>1.714</td>
</tr>
<tr>
<td>15/1/2004</td>
<td>0.457</td>
<td>1.709</td>
<td>1.365</td>
<td>2.743</td>
<td>1.705</td>
<td>1.709</td>
</tr>
<tr>
<td>16/1/2004</td>
<td>0.630</td>
<td>1.704</td>
<td>1.150</td>
<td>1.705</td>
<td>1.674</td>
<td>1.704</td>
</tr>
<tr>
<td>17/1/2004</td>
<td>1.240</td>
<td>1.698</td>
<td>0.764</td>
<td>0.370</td>
<td>1.657</td>
<td>1.698</td>
</tr>
</tbody>
</table>

* LT – Lower Troposphere; LS – Lower Stratosphere

The lower tropospheric and stratospheric AODs and their statistical tests for Sukhothai are essentially predicted. Results show that, AODs in the lower troposphere are in the range of 0.475-1.240 and 1.698-1.719 with the RMSE and MBD of 1.118 and 1.556 from the analysis of lidar data and the computational modeling, respectively. While, the lower stratospheric AODs are in the range of 1.657-1.712 with the RMSE and MBD of 0.539 and 0.015. In summary, the good correlation of AODs between the analysis of lidar data and the computational modeling for the lower stratosphere is revealed while existing similar trend-line for the lower troposphere. This suggests that the columnar AOD determination by computational methods are applicable and advantageous for the case of lacking of actual measurement and continuous AOD forecast. Still, a considerable improvement of the AOD retrieval remains interesting aspect for either parameter implementation or studying aerosol in deep details such as aerosol phase function and other optical properties in order to benefit the knowledge of aerosol in Thailand.

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