ความลึกเชิงแสงของละอองลอยที่ขึ้นกับความยาวคลื่นในปี 2546 ณ อำเภอศรีสำโรง จังหวัดสุโขทัย

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บทคัดย่อ

งานวิจัยนี้ได้แสดงขั้นตอนวิธีวิเคราะห์ความลึกเชิงแสงของละอองลอยในบรรยากาศ (AOD) ที่อำเภอศรีสำโรง จังหวัดสุโขทัย ด้วยข้อมูลจากเครื่องเรดิโอมิเตอร์แบบหมุนแถบเงาที่มีหลายตัวกรอง MFR-7 ที่ความยาวคลื่น 5 ค่า คือ 415 500 615 673 และ 870 นาโนเมตร โดยมีการพิจารณาค่าการกระเจิงแบบเรย์ลีและปริมาณตลอดคอลัมน์ในบรรยากาศของ NO_2 และโอโซน ในการศึกษาการเปลี่ยนแปลงของ AOD สำหรับช่วงละ 5 วันของเดือนมกราคมและมีนาคม 2546 ผลที่ได้แสดงให้เห็นว่าปริมาณตลอดคอลัมน์ในบรรยากาศของ NO_2 และ โอโซนอยู่ในช่วง 4-20 DU และ 250-370 DU ตามลำดับ ค่าพารามิเตอร์อังสตรอม β และ α โดยเฉลี่ยหาได้จากการปรับด้วยวิธีกำลังสองน้อยที่สุดอยู่ในช่วง 0.097-0.380 และ 0.910-1.442 ตามลำดับ ทั้งนี้ผลการวิเคราะห์ข้อมูลการสิ้นสุดที่ขึ้นกับความยาวคลื่นที่ได้นี้สามารถใช้เป็นข้อมูลละอองลอยในชั้นบรรยากาศเฉพาะบริเวณประเทศไทยซึ่งสามารถนำไปอ้างอิงในการศึกษาวิจัยด้านการแผ่รังสีและวิทยาศาสตร์บรรยากาศต่อไป

คำสำคัญ : การวิเคราะห์แบบแลงเลย์ / ท้องฟ้าโปร่ง / การแปรเปลี่ยนของอัตราการแพร่เชิงปกติ / อัตราการสิ้นสุด / ค่าพารามิเตอร์อังสตรอม

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Wavelength Dependence of Aerosol Optical Depth in 2003 at Sri Samrong District of Sukhothai Province in Thailand

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Abstract

The dependence on wavelength of the atmospheric Aerosol Optical Depth (AOD) at Sri Samrong district of Sukhothai province in Thailand were retrieved using multi-filter rotating shadowband radiometer (MFR-7) data at 5 distinct wavelengths of 415, 500, 615, 673 and 870 nm with the consideration of Rayleigh scattering and atmospheric column NO_2 and O_3 contents. Diurnal variations of AOD were accurately studied for 5-day-period in January and March 2003. Results show that the range of atmospheric column amounts of NO_2 and O_3 are 4-20 DU and 250-370 DU, respectively. The mean Ångström's parameters of β and α determined via least-square fitting method are in the range of 0.097-0.380 and 0.910-1.442, respectively. The obtainable results provide regional information of accurate spectral extinctions in Thailand which can advantageously provide the aerosol database for radiative forcing study and research in the future.

Keywords : Langley Analysis / Clear Sky / Normalized Diffuse Ratio Variability / Extinction Ratio / Ångström's Parameters

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1. Introduction

Aerosol climate effects are more complex and with large uncertainties [1] unlike the long-lived greenhouse gases. Atmospheric aerosol is a very important atmospheric trace component. Aerosols display substantial and temporal variations due to their short lifetimes within a week or less and various natural and anthropogenic emission sources. It deeply affects climate and land surface environment in both regional and global scales, whilst the tropospheric aerosols, e.g. carbonaceous (black carbon and organic carbon), sulfate and soil dust, much influence on human health. Tropospheric aerosol particles affect the Earth's radiation budget directly by scattering and absorbing solar and terrestrial radiation and indirectly by modifying the physical and radiative properties of clouds [2]. There is a rising awareness of the need to consider aerosol particles in climate modeling and global climate change. Aerosol Optical Depth (AOD) is a key parameter for the climate and environmental study, it is the extinction of the amount of incoming solar radiation caused by atmospheric aerosols which exists in various forms such as dust, fume and mist [3]. AODs at different wavelengths are related to their concentration, size distribution and aerosol optical properties, since aerosol scattering varies in complicated manner depending on size distribution and indices of refraction of the aerosol particles [4]. Until now many methods have been proposed to retrieve AODs, for instance several ground-based networks for aerosol optical properties were operated in East Asia [5] such as Aerosol Robotic Network (AERONET) sun/sky radiometer, SKYNET and the Asian Dust Lidar Network (using micro-pulse lidar: MPL). From AERONET database, AODs at 440 nm wavelength of Thailand in 2003 were overestimated, because AERONET retrieval algorithm neglected the contribution of NO2 absorption to the retrieved optical depths [6]. And Shaw [7] confirmed that significant errors in AOD estimates could be arisen if the absorption by NO_2 was neglected. In 1999, Jacovides et al. [8] retrieved aerosol spectral optical depths of the polluted Athens atmosphere using ground-based spectral solar extinction data. They indicated the absorptions due to NO_2 and O_3 by UV spectral signature at specific wave band. Later in 2010 [9], the retrieval algorithm of total atmospheric column NO_2 amounts at 415 nm wavelength in 2003 at Sri Samrong district of Sukhothai province (17° 0′ 21″ N, 99° 49′ 35″ E), Thailand were carried out by using MFR-7 as ground-based measurement of solar irradiance data, and automatically AOD value was obtained. Since the observation station is located in countryside where has sunny continental climate with a dominance of small soil dust particles. Such topography was considered for representing a characteristic of general unpolluted rural area in Thailand.

In this work, we implemented the result at 415 nm wavelength [9] by analyzing other 4 wavelength dataset in an attempt to improve a technique with the minimization of possible errors for good estimates of spectral AOD. Therefore the results of accurate AODs and their spectral dependences of 5 distinct wavelengths of 415, 500, 615, 673 and 870 nm, being subtracted by optical depths of NO_2, O_3 and Rayleigh scattering retrieved from MFR-7 measurements over the atmosphere of Sri Samrong region are shown to support the aerosols study in Thailand. This also yields the regional information of spectral extinction and systematic retrieval of aerosol optical properties as by-product. Atmospheric column NO_2, O_3 optical depths, AOD and instrumental calibration constant retrievals were

done by Langley analysis from cloudless-day data at those 5 wavelengths. An identification of measured data of clear sky periods was simply provided by applying the normalized diffuse ratio variability test [10] together with an elimination of both early morning and late evening periods when solar zenith angles are greater than 80° in average. Finally the Ångström exponent (α) and the Ångström turbidity coefficient (β) were additionally calculated with the available AOD spectra in order to examine aerosol particles in deep detail such as size distribution.

2. Instrumentation and Observation

MFR-7 [11] is a field instrumental that measures global or total, direct and diffuse components of solar irradiances with up to 7 wavelengths. It is equipped with one broadband channel and six narrowband channels with 10 nm FWHM centered on 415, 500, 615, 673, 870 and 940 nm. A microprocessor-controlled shadowband alternately shades and exposes an instrumental diffuser, enabling the system to measure all three irradiance components by one detector. The band is rotated to measure total irradiance and the diffuse horizontal irradiance is measured when the sun is completely blocked. In order to obtain the direct normal component, total irradiance is subtracted from corrected diffuse component and divided by the cosine of solar zenith angle.

From the typical optical depths of atmospheric constituents in the MFR-7 spectra [6], it was found that aerosols and Rayleigh scatterings contribute atmospheric extinction in all channels. The dominant gaseous absorbers within the first five MFR-7 channels are NO_2 (at 415 and 500 nm) and O_3 (at 500, 615 and 673 nm), although water vapor absorption contributes dominant attenuation to the solar radiation at 940 nm. And aerosol extinction of the direct solar radiation of various wavelengths in the range < 732 nm is usually larger than ones > 732 nm [10]. With all these properties, AOD retrievals were firstly done from the measured data at 870 nm-channel since the extinction source is primary aerosol with little contribution from gaseous absorbers, and then daily AOD, NO_2 and O_3 optical depths were preferably determined from the measured data at the rest of 4 different wavelength-channels (415, 500, 615 and 673 nm), while the measured data at 940 nm-channel was unsuitable for AOD study according to the reason mentioned earlier.

3. Retrieval Algorithm

In order to optimize the good measured data for AOD retrieval, the clear sky measured data should be investigated and performed. Therefore, the measured data was initially normalized by a power law function of the cosine of solar zenith angle [10]:

$$D_n = D_{\downarrow} / \mu^{-0.5} \quad \textbf{(1)}$$

where $D_{\downarrow}$ is the downwelling diffuse ratio shortwave (the diffuse shortwave irradiance divided by the total shortwave irradiance), D_n refers to the normalized diffuse ratio shortwave and μ is the cosine of solar zenith angle. The deviation data in terms of normalized irradiance were excluded from the clear sky periods with an assumption of the presence of clouds and temporal variations in haze or subvisual cirrus. The valid minimum number of identified clear sky measurements is set at 120, to assure the sufficient data for obtaining good results. Consequently a process of retrieval algorithm for the measured direct normal solar irradiance data of daily clear sky periods is able to perform properly for clear day at all 5 wavelength-channels. Due

to the high absorption coefficients of NO_2 and O_3 respectively at 415 nm and 615 nm, and very small absorption cross sections of NO_2 above 600 nm, therefore we neglected the optical depth of NO_2 in this wavelength-region, and the algorithm for retrievals of AODs, NO_2 and O_3 optical depths with a simultaneous determination of the instrumental calibration constants via Langley analysis of direct normal solar irradiance data set were done via the following equations [6]:

$$\tilde{\tau}'_{870}\, m = \tau_a\, m + Inc \quad \textbf{(2)}$$

$$\tilde{\tau}'_{415}\, m = (q\tau_a + \tau_{NO_2(415)})m + Inc_{415} \quad \textbf{(3)}$$

$$\tilde{\tau}'_{615}\, m = (q'\tau_a + \tau_{O_3(615)})m + Inc_{615} \quad \textbf{(4)}$$

where $\tilde{\tau}'_{870}$, $\tilde{\tau}'_{415}$ and $\tilde{\tau}'_{615}$ stand for the total measured optical depth with the removal of Rayleigh component in each channel respectively, q and q′ are extinction ratio normalized to the 870 nm-channel of 415 and 615 nm-channel respectively, and the subscript a, NO_2 and O_3 are aerosol, NO_2 and O_3. Here c, c_{415} and c_{615} are instrumental calibration constants with respect to 870, 415 and 615 nm-channel and m is the solar air mass relative to unit air mass in zenith direction with a common air mass formula [11]:

$$m = \left[\cos(z) + 0.50572(96.07995\text{-}z)^{-1.6364}\right]^{-1} \quad \textbf{(5)}$$

here z is solar zenith angle in degree. Since the measured direct normal solar irradiance can be represented in the non-calibrated form :

$$I = I_0 \exp(-\tilde{\tau}\, m) \quad \textbf{(6)}$$

where I_0 is the top of atmosphere solar intensity, while ln I_0 has been obtained via Langley plot of ln I versus m (not greater 3). And then $\tilde{\tau}\, m$ is determined which is equal to (ln I_0 – ln I). In order to remove Rayleigh scattering component from the total measured optical depth ($\tilde{\tau}$), we use $\tau_{Rayleigh}$ as reported in Table 1 which were calculated by a relative equation [11]:

$$\tau_{Rayleigh} = 0.008569\lambda^{-4}(1 + 0.0113\lambda^{-2} + 0.00013\lambda^{-4})\frac{P}{P_0} \quad \textbf{(7)}$$

where λ is the wavelength in micrometers, P (~1008.80 mbar) is the mean value of site pressure relative to sea level pressure P_0 (1013.25 mbar).

Table 1 Wavelength dependences of Rayleigh Optical Depth ($\tau_{Rayleigh}$), NO_2 absorption coefficient (σ_{NO_2}) and O_3 absorption coefficient (σ_{O_3})

λ (nm)	$\tau_{Rayleigh}$	σ_{NO_2} x 10^{-19} (cm^2 / molecule)	σ_{O_3} x 10^{-22} (cm^2 / molecule)
415	0.3077	6.03	0.0014
500	0.1409	3.00	11.98
615	0.0626	-	43.03
670	0.0432	-	15.81
870	0.0151	-	0.005

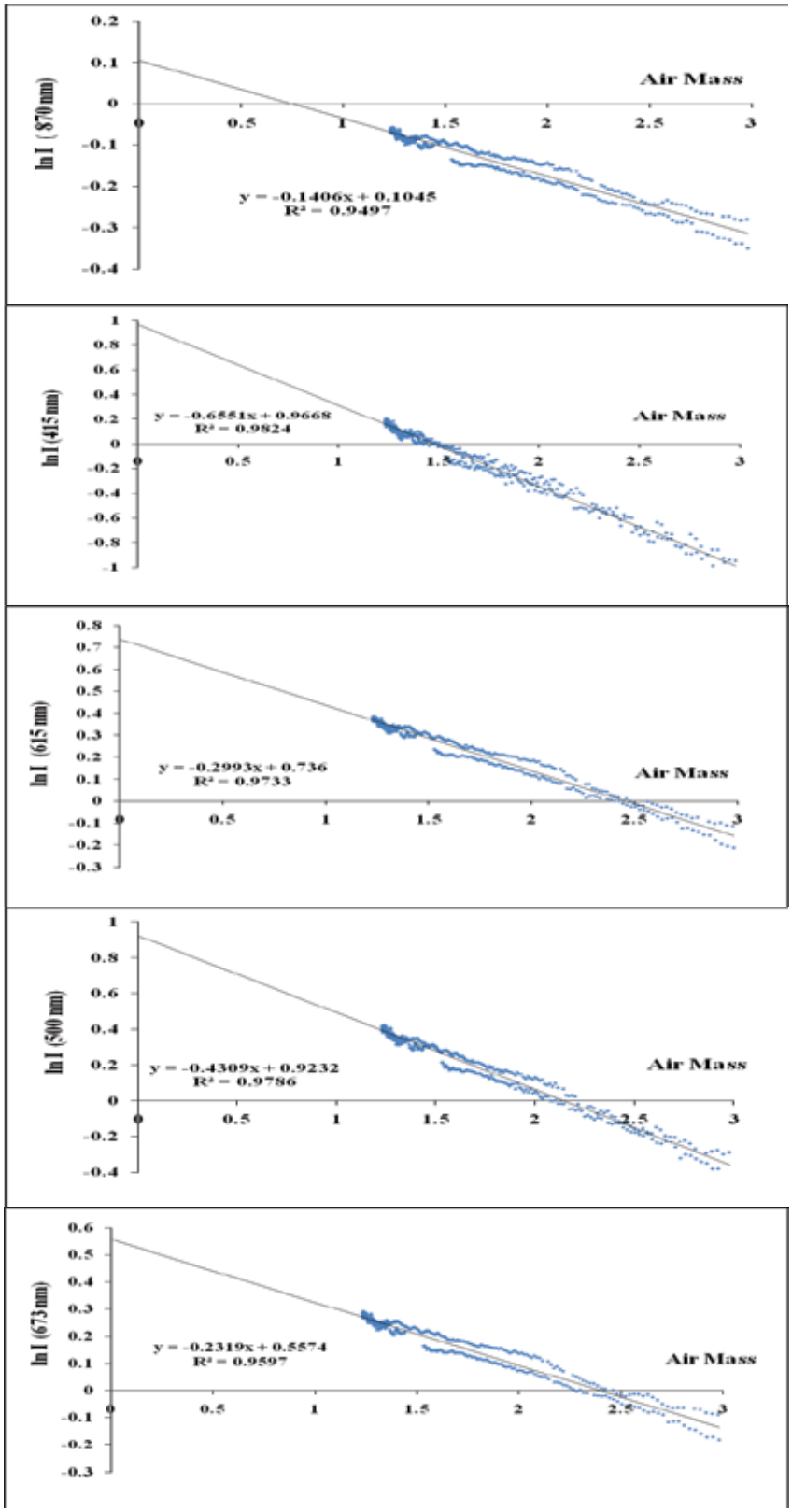
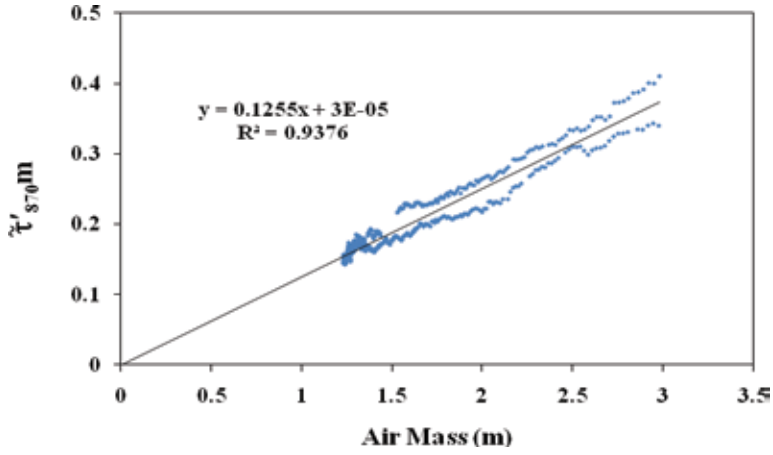
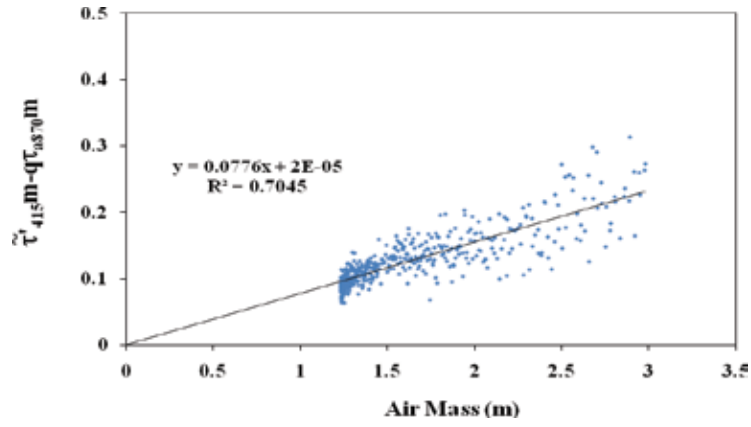


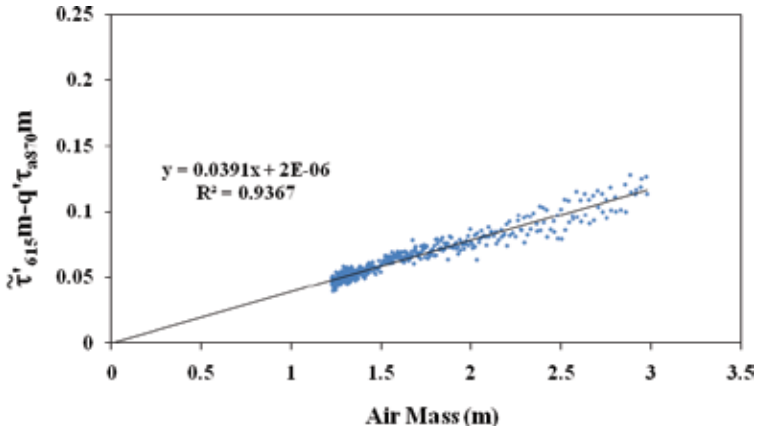
Fig. 1 Langley plots of identified clear sky period on 26 January 2003 for non-calibrated Eq. (6) of 5 distinct wavelengths.



(2a)

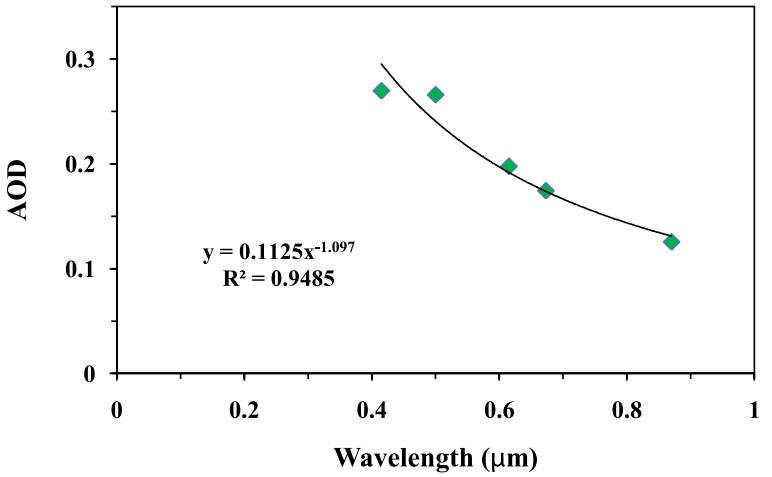


(2b)

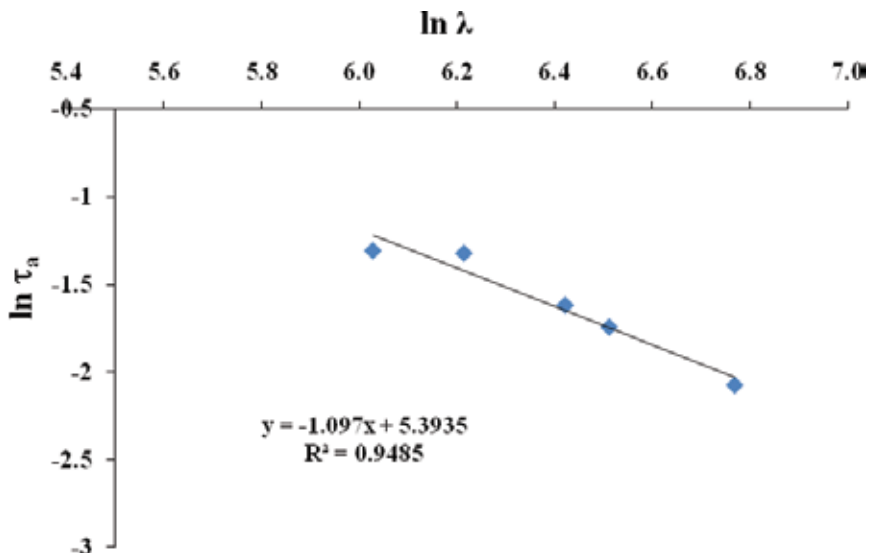


(2c)

Fig. 2 Regression plots on 26 January 2003 (2a) of Eqn 2 for retrieval of optical depth in 870 nm-channel, (2b) of Eqn 3 for retrieval of optical depth in 415 nm-channel with q = 2.15, and (2c) of Eqn 4 for retrieval of optical depth in 615 nm-channel with q′ = 1.575.



(3a)



(3b)

Fig. 3 AOD spectrum on 26 January 2003 at the observation site is shown in (3a) and (3b) is the log-log fit of Eq. (12) on the same day. Results expressed that Ångström exponent $\alpha = 1.097$ and the Ångström turbidity coefficient $\beta = 0.1125$.

Whenever τ_{NO_2} and τ_{O_3} are available at 415 and 615 nm-channel respectively (whereas NO_2 and O_3 are the major absorbers), the NO_2 and O_3 column amounts are determined in Dobson units (1 DU = 10^{-3} atm-cm = 2.687 x 10^{16} molecule.cm^{-2}) with the effective spectral absorption coefficients (σ) and the relations [12] of

$$\tau_{NO_2}(\lambda) = \sigma_{NO_2}(\lambda)\, N_{NO_2} \quad (8)$$

$$\tau_{O_3}(\lambda) = \sigma_{O_3}(\lambda)\, N_{O_3} \quad (9)$$

here N_{NO_2} and N_{O_3} are the total column NO_2 and O_3 amounts respectively.

By means of these retrievals, τ_{NO_2} and τ_{O_3} at other wavelengths are provided, and then the aerosol optical depths can be obtained using following equations:

$$\tilde{\tau}'_{500}\, m = (\tau_{a(500)} + \tau_{NO_2(500)} + \tau_{O_3(500)})m + Inc_{500} \quad (10)$$

$$\tilde{\tau}'_{673}\, m = (\tau_{a(673)} + \tau_{O_3(673)})m + Inc_{673} \quad (11)$$

With the above assumptions, firstly τ_a and c for 870 nm-channel were found by a least square fitting method with the plot of $\tilde{\tau}'_{870}$ versus m. Secondly the parameterization procedure of q value from Eqn 3 was optimized by means of regression technique with the known values of τ_a , $\tau_{NO_2(415)}$ and then was subsequently available (with the square of correlation coefficient $R^2 > 0.7$). Similar retrievals were done for the parameter q′ and in Eq. (4). Next $\tau_{NO_2(500)}$ and $\tau_{O_3(500)}$ were determined and finally $\tau_{a(500)}$ and $\tau_{a(673)}$ were retrieved from Eq. (10) and (11). Hence the spectral AOD was obtained.

The Ångström formula [12] for approximating aerosol extinction is given by:

$$\tau_a(\lambda) = \beta\lambda^{-\alpha} \quad (12)$$

where λ is the wavelength in μm, β is known as the turbidity, and α is Ångström exponent which relates to the aerosol particle type and size. The small value of α signified large size particle [13], $\alpha \sim 0$ corresponded to large dust particle and so did $\alpha \sim 2$ for fine smoke aerosol. The range of α = 1.54–1.66 [14] normally found in urban area with air pollution of soot aerosols from heavy vehicular traffic and industrial activity. Moreover in 2002, α were found in a range of 0.54–1.96 with the mean of summer month values of α = 1.45 in Estonia (marshy lowland terrain in temperate climate zone), including total summer mean β of 0.091 [15].

4. Result and Discussion

Fig. 1 shows Langley plot of Eq. (6) to obtain the total atmospheric column optical depth ($\tilde{\tau}$) and ln I_0 for each 5 wavelengths. By using regression techniques with Eq. (3) and (4) (Fig. 2), all unknowns especially q and q′ can be found with high correlation in the range greater than 0.7-1.0. Diurnal results of total atmospheric NO_2 and O_3 amounts are summarized in Table 2, and Table 3 shows the values of atmospheric AOD for 5 wavelengths. Also Fig. 3b is the log-log fit of equation:

$$\ln \tau_a(\lambda) = -\alpha \ln \lambda + \ln \beta \quad (13)$$

whereas the slope yields Ångström exponent and the intercept provides Ångström turbidity coefficient β. With the high values of R^2, an efficiency of this technique is validated. Fig. 4 shows daily plots of the mean Ångström exponents $\alpha_{415\text{-}870}$ of 0.910-1.392 in winter (January) and 1.055-1.442 in summer (March), with the mean

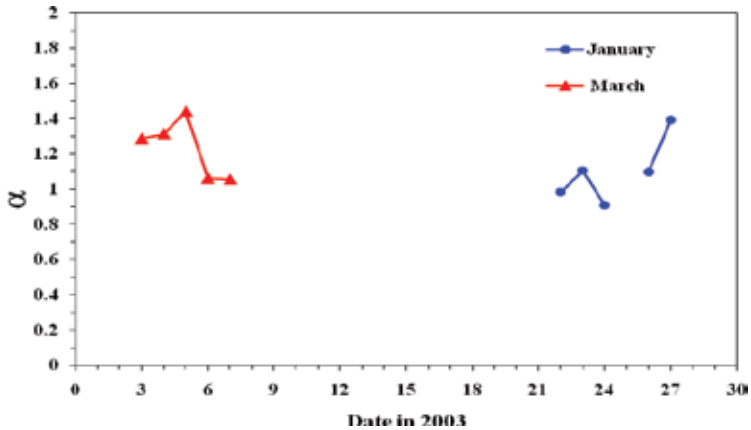
Ångström turbidity coefficients β of 0.0968-0.1826 and 0.2201-0.3800, respectively. Daily spectral AODs were plotted for 5 cloudless-day observation in January and March, 2003, as shown in Fig. 5. Lastly, diurnal variations of total column NO_2 and O_3 amounts for winter and summer were plotted in Fig. 6.

Table 2 Diurnal total atmospheric NO_2 and O_3 amounts

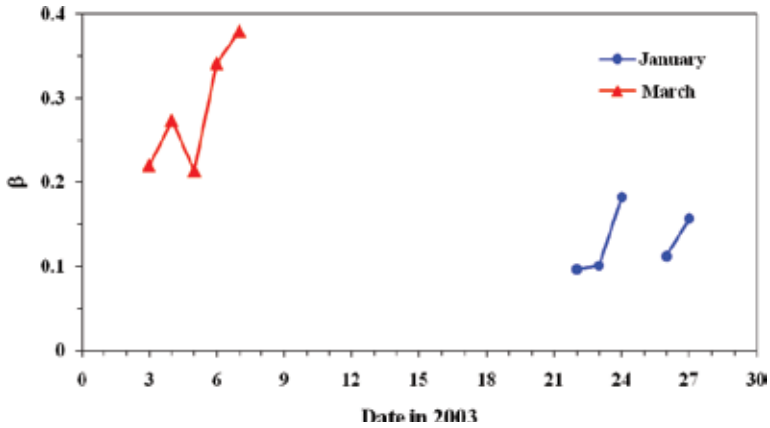
Date in 2003	**NO_2** (DU)	**O_3** (DU)
22nd Jan	3.70	249.9
23rd Jan	5.02	323.5
24th Jan	6.38	307.9
26th Jan	4.79	338.2
27th Jan	4.88	308.9
3rd Mar	18.63	344.2
4th Mar	16.21	328.7
5th Mar	11.10	298.4
6th Mar	14.89	370.2
7th Mar	19.54	326.9

Table 3 Daily values of atmospheric AODs for 5 wavelengths, and the mean Ångström exponent (α) for winter (January) and summer (March) with the mean Ångström turbidity coefficient (β).

Date in 2003	**AODs for 5 wavelengths**					α	β
	415 nm	**500** nm	**615** nm	**673** nm	**870** nm		
22nd Jan	0.2124	0.2106	0.1593	0.1493	0.1062	0.983	0.0968
23rd Jan	0.2483	0.2377	0.1806	0.1617	0.1129	1.105	0.1013
24th Jan	0.3770	0.3706	0.2926	0.2655	0.1984	0.910	0.1826
26th Jan	0.2698	0.2660	0.1977	0.1744	0.1255	1.097	0.1125
27th Jan	0.4638	0.5084	0.3014	0.2656	0.1855	1.392	0.1568
3rd Mar	0.6311	0.5659	0.4455	0.3705	0.2475	1.288	0.2201
4th Mar	0.7836	0.7456	0.5531	0.4706	0.3073	1.312	0.2740
5th Mar	0.6869	0.6695	0.4325	0.3740	0.2544	1.442	0.2141
6th Mar	0.8012	0.7815	0.5913	0.5191	0.3815	1.062	0.3418
7th Mar	0.8730	0.8668	0.6651	0.5851	0.4157	1.055	0.3800



(4a)



(4b)

Fig. 4 Diurnal variations of 4a) the mean Ångström Exponent (α) and 4b) the mean Ångström turbidity coefficient (β) for winter (January) and summer (March) 2003.

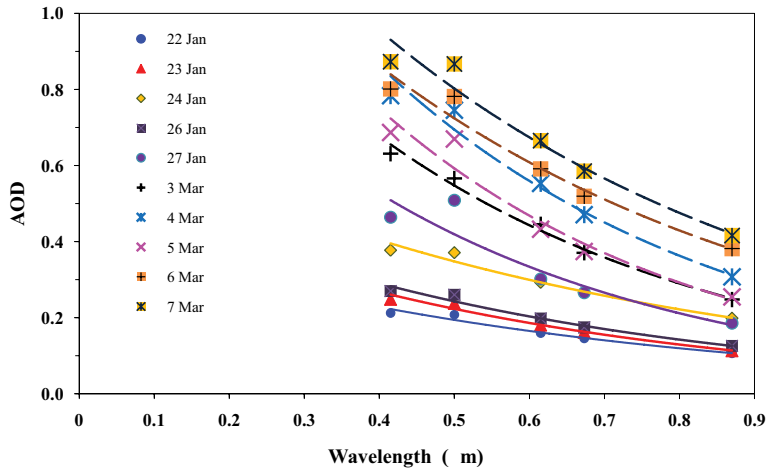
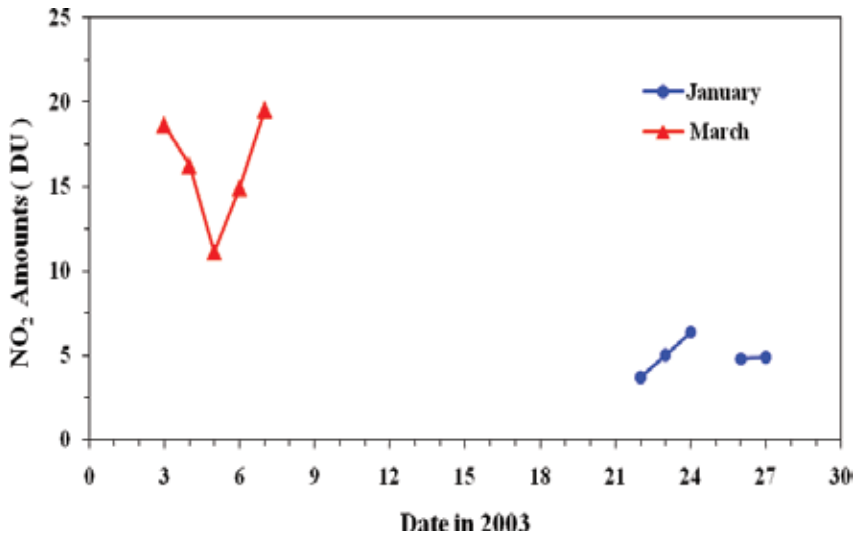
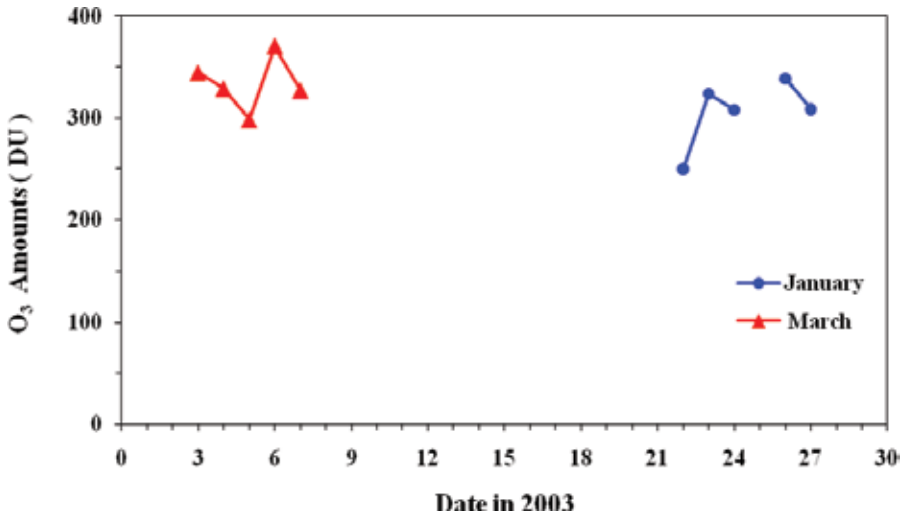


Fig. 5 Variations of daily spectral AODs for each 5 day in January and March 2003.



(6a)



(6b)

Fig. 6 Plots of diurnal variations of (6a) total column NO_2 amounts and (6b) total column O_3 amounts.

To monitor the reliabilities of spectral AOD results, the atmospheric column NO_2 and O_3 amounts were preferably verified. Therefore we compared our results with the O_3 profile at the South of Thailand using Brewer spectrophotometer and UMKEHR technique for their determinations. The approximated annual average of 4.70 x 10^{12} molecules. cm^{-3} was found in 2003 [16]. Besides, Ozone vertical column collected from Dobson and TOMS instrument of ~250 and ~270 DU respectively (~in the order of 10^{18} molecules.cm^{-2} during 2000-2003), but differential slant column densities

of O_3 found in an order of 10^{20} molecule.cm^{-2} during 2000 at Pune, India [17] were reported in 2007. The research concluded that column amounts of O_3 vary with season and altitude. Also, there existed a research [18] reporting that the Stratospheric Aerosol and Gas Experiment (SAGE II) measurement of ozone tended to slightly overestimate ozone at high altitude and systematically underestimated it in troposphere by approximately 30% in the regions between 8 km altitude and 2 km below the tropopause. With all results of column ozone stated in three references [16-18], the vertical column ozone amounts as seen in Fig. 6b seems to be slightly overestimated, and seasonal invariance. Reasons for discrepancies between the obtainable result in this study and the other estimates are yet unclear, however it is scarcely understood whether or not the results are overestimated. Nevertheless, it is well known that ozone has small absorption in visible wavelengths and the overestimation of retrieved ozone optical depths in 615 nm-channel may cause by haze absorption which is often associated with high level enhancement of tropospheric ozone [19].

From the results of atmospheric column NO_2 amount retrievals, in Fig 6a and Table 2, they suggest that our results are in better agreement with the neighbors in tropical Asian region (at Pune, India) [17] than ones in our previous work [9]. This probably relate to more accurate techniques of analysis, namely the limit of air mass (m) used in this work is less than 3 instead of 2.5. Furthermore, the regression analysis of each AOD spectrum is very helpful for an adjustment of q value (extinction ratio normalized to the 870 nm-channel of 415 nm-channel), which is a significant parameter for optimizing both results of atmospheric column NO_2 amounts and AODs.

With all information of the comparison of O_3 and NO_2 amounts, it supports that the obtainable AOD spectra are accurate and reliable.

5. Conclusions

Since the validation of column NO_2 content retrievals were monitored [9], this study was the next step for retrieval a better estimate AODs in a wavelength range from 415 to 870 nm, 5 distinct wavelengths in total. As a result, our focus was on the best fit of obtained AOD spectra with automatic retrieval of reliable values of total column NO_2 and O_3 amounts and also the obtained spectral AODs were derived independently without using another supported data from other different instruments.

Results reveal the feasibility of retrieval algorithms of diurnal variation atmospheric aerosols as well as total column NO_2 and O_3 amounts in the research location from measured MFR-7 data. It is observed that total column NO_2 amounts are higher in summer than ones in winter, whilst total column O_3 amounts are seasonal invariance. In case of the obtained AOD spectra, behavior of the retrieved AODs at shorter wavelengths is higher than ones at longer wavelengths; it implies that extinction efficiencies decrease with the wavelength. Importantly all obtainable instrumental calibration constants express that the well-calibrated instrument was rather stable with regard to the quality of the measurements.

The available results indicate the mean wavelength exponent $\alpha_{415\text{-}870}$ of 0.910-1.442 are rather low, compared to the mean value of α found in the cited references [13-15] while higher aerosol loadings were observed (implicitly by high β) during summer (March) as seen in Fig. 4b. Besides, the higher AOD values observed during summer would be contributed by anthropogenic activities

in aerosol loading of black carbon from local biomass burning due to routine behavior of Thai farmers. Such phenomenon is evident from seasonal variations of the atmospheric turbidity because high turbidity days were observed in summer (March). Therefore, values of β in the hot season are higher than ones in winter, since high temperature contributes an increment of heat convection with uplifts mixture of biomass-burning aerosols with soil dust particles from ground to the atmosphere. The results (the spectral extinction behavior of aerosols in Thailand) can subsequently be represented a typical unpolluted rural-background (without urban pollution) aerosol conditions in Thailand. However, the consequence of anthropogenic activities in aerosol loading might be observed in local climate and should be helpful for further regional AOD modeling studies in future.

In summary, we would like to pinpoint that the small uncertainties of our results are partly based on uncertainties of Rayleigh scattering optical depth (using constant average value of site pressure P), and absorption coefficients of NO_2 and O_3 for distinct wavelengths. In addition, errors in NO_2 and O_3 amounts can result in variations of AODs for all 5 wavelengths. However, the good fits with high values of R^2 in Fig. 2 and 3 shows the efficiency of this technique for spectral aerosol optical depth analysis from MFR-7 data. It is very useful for provision of the accurate spectral dependence of aerosol single scattering albedo in future, suggesting the progressive studies of radiative forcing in Thailand.

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