Utilization of O_4 slant column density to derive aerosol layer height from a spaceborne UV-Visible hyperspectral sensor: Sensitivity and case study

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Abstract

The sensitivities of oxygen-dimer (O_4) slant column densities (SCDs) to changes in aerosol layer height are investigated using the simulated radiances by a radiative transfer model, the Linearlized pseudo-spherical vector discrete ordinate radiative transfer (VLIDORT), and the Differential Optical Absorption Spectroscopy (DOAS) technique. The sensitivities of the O4 index (O4I), which is defined as dividing O_4 SCD by 10^{40} molecules^{2}cm^{-5}, to aerosol types and optical properties are also evaluated and compared. Among the O_4 absorption bands at 340, 360, 380, and 477 nm, the O_4 absorption band at 477 nm is found to be the most suitable to retrieve the aerosol effective height. However, the O4I at 477 nm is significantly influenced not only by the aerosol layer effective height but also by aerosol vertical profiles, optical properties including single scattering albedo (SSA), aerosol optical depth (AOD), particle size, and surface albedo. Overall, the error of the retrieved aerosol effective height is estimated to be 1276, 846, and 739 m for dust, non-absorbing, and absorbing aerosol, respectively, assuming knowledge on the aerosol vertical distribution shape. Using radiance data from the Ozone Monitoring Instrument (OMI), a new algorithm is developed to derive the aerosol effective height over East Asia after the determination of the aerosol type and AOD from the MODerate resolution Imaging Spectroradiometer (MODIS). About 80% of retrieved aerosol effective heights are within the error range of 1 km compared to those obtained from the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) measurements on thick aerosol layer cases.

1. Introduction

Aerosol is one of the key atmospheric constituents in understanding climate changes with its effects on direct and diffuse solar radiation (e.g., Haywood and Shine, 1995; Kaufman *et al.*, 2002), and plays an important role in air quality near the surface (e.g., Watson *et al.*, 1994; Prospero, 1999). For these reasons, observations from satellite remote sensing have been carried out to investigate aerosol properties at regional and global scale, including aerosol optical depth (AOD) (e.g., Curier *et al.*, 2008; Levy *et al.*, 2007; Torres *et al.*, 2007; Ahn *et al.*, 2014; Veefkind *et al.*, 1999; Zhang *et al.*, 2011), fine mode fraction (FMF) or Angstrom Exponent (AE) (e.g., Jones and Christopher, 2007; Lee *et al.*, 2010; Nakajima and Higurashi, 1998; Remer *et al.*, 2008), single scattering albedo (SSA) (e.g., Dubovik *et al.*, 2002; Levy *et al.*, 2007; Jeong and Hsu, 2008; Torres *et al.*, 1998, 2005, 2007; Jethva *et al.*, 2014), and aerosol types (e.g., Higurashi and Nakajima, 2002; Kim *et al.*, 2007; Lee *et al.*, 2010). These information were further utilized to estimate radiative forcing of aerosol (e.g., Christopher *et al.*, 2006; Chung *et al.*, 2005; Chou *et al.*, 2002), to understand the mechanism of the changes to the cloud formation (Twomey *et al.*, 1984; Albrecht, 1989; Jones *et al.*, 1994), and to monitor air quality (e.g., Wang and Christopher, 2003; Hutchison *et al.*, 2005).

Vertical profiles of atmospheric aerosols are affected by processes of formation, transport and deposition, and vary for different aerosol types over East Asia (Shimizu *et al.*, 2004). Labonne *et al.* (2007) also reported that the layer top height of biomass burning aerosol ranged from 1.5 to 7 km in the wild fire regions. The information on the aerosol layer height is important, because the variation of the aerosol vertical distribution affects radiative process in the atmosphere near the surface and trace gas retrieval for air mass factor calculation. Uncertainty in aerosol layer height also affects the accuracy of AOD and SSA retrieval algorithms that use near UV observations (Torres *et al.*, 1998; Torres *et al.*, 2007; Jethva *et al.*, 2014) and complicates the interpretation of the Aerosol Index (AI), a qualitative parameter commonly used to detect absorbing aerosols (Herman *et al.*, 1997; Torres *et al.*, 1998). In addition, there have been difficulties to estimate surface concentration of aerosol from AODs, because the information on aerosol vertical distribution is not readily available and even hard to predict from the state-of-the-art models due to its large variability. Although the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) has been successful and provided vertical profiles of aerosols, its spatial coverage was very limited with its measurement characteristics (Omar *et al.*, 2009). Liu *et al.* (2005) showed that the Particulate Matter (PM) concentration estimated by the AOD from satellite observation accounted for only 48% of the measured surface PM, although their study reflected variations of the aerosol types and its hygroscopic growth in the algorithms. One of the essential factors to consider in estimating PM from AOD is the vertical structure of aerosols (e.g. Chu, 2006; Seo *et al.*, 2015). Therefore, conventional aerosol products would benefit significantly with the development of robust algorithm to retrieve aerosol height using satellite data.

The Differential Optical Absorption Spectroscopy (DOAS) technique has been used widely to retrieve trace gas concentration both from ground-based (e.g., Platt, 1994; Platt and Stutz, 2008) and space-borne (e.g., Wagner *et al.*, 2007; Wagner *et al.*, 2010) measurements. After the work of Platt (1994) to retrieve trace gas concentration by using DOAS, Wagner *et al.* (2004) suggested to derive atmospheric aerosol information from O_4 measurement by using Multi Axis Differential Optical Absorption

Spectroscopy (MAX-DOAS). Friess et al. (2006) analyzed the model studies to calculate the achievable precision of the aerosol optical depth and vertical profile. In addition, several studies (e.g., Irie *et al.*, 2009 and 2011; Lee *et al.*, 2009 and 2011; Clemer *et al.*, 2010; Li *et al.*, 2010) provided aerosol profiles from ground-based hyperspectral measurements in UV and visible wavelength ranges on several ground sites. Wagner *et al.* (2010) investigated the sensitivity of various factors to the aerosol layer height using the data obtained from the SCanning Imaging Absorption SpectroMeter for Atmospheric ChartographY (SCIAMACHY) on ENVISAT. The sensitivity of the Ring effect and the absorption by oxygen molecule (O_2) and its dimer (O_4) calculated by DOAS method were examined to estimate aerosol properties including the layer height. Kokhanovsky and Rozanov (2010) estimated dust altitudes using the O_2-A band between 760 and 765 nm after the determination of the dust optical depth. In addition, several previous studies are also investigated estimation methods for aerosol height information by using hyperspectral measurement in visible (e.g., Dubuisson *et al.*, 2009; Koppers and Murtagh, 1997; Sanders and de Haan, 2013; Sanghavi *et al.*, 2012; Wang *et al.*, 2012). Because the surface signal is significantly smaller than the aerosol signal in the near UV, these wavelength regions are useful to derive aerosol height information from space borne measurements.

For OMI measurement, the O_4 band at 477 nm has been widely applied to estimate cloud information (e.g., Accarreta *et al.*, 2004; Sneep *et al.*, 2008). Especially, the cloud information retrieved by O_4 band at 477 nm was used to analyze air mass factor (AMF) with the consideration of aerosol optical effects for the NO_2 column retrieval (e.g., Castellanos *et al.*, 2015, Chimot *et al.*, 2015; Lin *et al.*, 2014; Lin *et al.*, 2015). Although O_4 absorption band around 477 nm varies also due to cloud existence, it can be also used for the aerosol optical parameter estimation. Veihelmann *et al.* (2007) introduced that the 477 nm channel, a major O_4 band, significantly adds degree of freedom for aerosol retrieval by using principal component analysis, and Dirksen *et al.* (2009) adopts the pressure information obtained from OMI O_4 band to identify a plume height for aerosol transport cases.

In this study, the sensitivities of the O_4 bands at 340, 360, 380, and 477 nm to changes in aerosol layer height and its optical properties are estimated using simulated hyperspectral radiances, differently from the previous studies using the O_2-A band observation (e.g., Kokhanovsky and Rozanov, 2010). We proposed an improved DOAS algorithm for the O_4 absorption bands to retrieve aerosol height information from the O_4 slant column densities (SCDs) based on the sensitivity studies. This new algorithm is applied to the O_4 SCD from the Ozone Monitoring Instrument (OMI) to retrieve the aerosol effective height (AEH) for a real case over East Asia, including error estimates.

2. Methods

In general, scattering by aerosol at low altitudes leads to an increase in the length of the average light path (enhancement effect), while those at high altitudes causes a decrease in the length of the average light path (shielding effect) (Wagner *et al.*, 2010). These two opposing effects change the estimated O_4 SCD values. Furthermore, the measured O_4 SCD is a function of wavelength, because the absorption and scattering by atmospheric molecules and aerosols have spectral dependence. Therefore, radiative transfer calculations are carried out to estimate the sensitivity of the O_4 SCD with respect to the change of atmospheric conditions. Details of the radiative transfer model (RTM) and input parameters to simulate radiance are discussed in section 2.1.

Analytical method of the DOAS to estimate the O_4 is described in section 2.2.

2.1. Simulation of hyperspectral radiance

Figure 1 shows the flowchart of the method to estimate the O_4 SCD from the simulated radiance. Because the magnitude of the O_4 SCD values is too large to express the sensitivity results, this paper defines the O_4 index (O4I) which divides O_4 SCD by 10^{40} molecules^{2}cm^{-5}. In order to investigate the sensitivities of the O4I at several bands in UV and visible wavelengths with respect to various aerosol properties, including AEHs, aerosol amounts and aerosol types, the hyperspectral radiance is simulated using the Linearlized pseudo-spherical vector discrete ordinate radiative transfer (VLIDORT) model (Spurr, 2006). The VLIDORT model is based on the linearized discrete ordinate radiative transfer model (LIDORT) (Spurr *et al.*, 2001; Spurr, 2002). This RTM is suitable for the off-nadir satellite viewing geometry of passive sensors since this model adopts the spherically curved atmosphere to reflect the pseudo-spherical direct-beam attenuation effect (Spurr *et al.*, 2001). The model calculates the monochromatic radiance ranging from 300 to 500 nm with a spectral resolution of 0.1 nm. The radiance spectrum is calculated with a 0.2 nm sampling resolution applying a slit response function (SRF) given by a normalized Gaussian distribution with 0.6 nm as the full-width half maximum (FWHM).

2.1.1. Aerosol properties

The aerosol input parameters for the RTM are important in simulating the radiance spectra because aerosol optical properties determine scattering and absorption characteristics. The data from the Optical Properties of Aerosol and Cloud (OPAC) package (Hess *et al.*, 1998) are used as aerosol parameters, which includes the spectral complex refractive indices and size distribution of aerosols, to calculate SSA and phase function through the Mie calculations. The information of the aerosol parameters is not available at the UV wavelengths, since the AERONET observation provides the information of those aerosol parameters in the visible.

In terms of the aerosol types, water soluble (WASO), mineral dust (MITR), and continental polluted (COPO) model are selected to simulate non-absorbing aerosol, mineral dust, and absorbing anthropogenic aerosol, respectively. The COPO is combined type including both soot and WASO, which represents the pure black-carbon and non-absorbing aerosols, respectively. The mixture of these two types, adequately describes the fine mode aerosol from anthropogenic pollution. The SSA is the largest for WASO and the smallest for COPO. In order to account for hygroscopic growth, the default relative humidity is assumed to be 80 % (c.f., Holzer-Popp and Schroedter-Homscheidt, 2004).

2.1.2. Aerosol vertical distribution

In this present study, 'aerosol height' refers to aerosol effective height (AEH), defined as the peak height in Gaussian distribution. According to Hayasaka *et al.* (2007), however, the aerosol extinction coefficient was found to exponentially decrease with altitude over East Asia based on the ground-based LIDAR observation data during the Atmospheric Brown Clouds-East Asia Regional Experiment 2005 (ABC-EAREX 2005) campaign. Previous studies used the exponentially decreasing pattern with altitude to represent the aerosol vertical profiles (e.g. Hayasaka *et al.*, 2007; Li *et al.*, 2010), and reported that aerosol is present within 5 km in altitude for most of the cases (e.g. Sasano, 1996; Chiang *et al.*, 2007). On the other hands, the aerosol vertical distribution does not always follow exponential profile. For the long-range transported aerosol such as dust cases, the aerosol layer profile is quite different than exponential profile and occasionally transported to well above the boundary layer (e.g., Reid *et al*., 2002; Johnson *et al.*, 2008). The peak height of aerosol extinction profile in long-range transport cases was reported to be located between 1 and 3 km during the Dust and Biomass-burning Aerosol Experiment (DABEX) campaign (Johnson *et al.*, 2008). From these previous studies, standard aerosol vertical profile is difficult to determine. For algorithm development, previous studies assumed that the vertical distribution is Gaussian function defined by peak height and half width as representative parameters (Torres *et al.*, 1998; Torres *et al.*, 2005). To supplement the simplicity of assumption for aerosol vertical distribution, aerosol vertical distributions are assumed to be quasi-Gaussian generalized distribution function (GDF), which is Gaussian distribution with dependence on aerosol peak height, width, and layer top and bottom height. Details of GDF can be found in Spurr and Christi (2014) and Yang *et al.* (2010). In this study, AEH ranges from 1 to 5 km with 1 km width as 1-sigma for the RTM simulation.

2.1.3. Atmospheric gases

The vertical distribution of the O_4 number density, which is used to calculate its SCD from the RTM, is assumed to be the square of the O_2 number density in each layer (Hermans *et al.*, 2003). Thus, the total number of the O_4 column density from surface to TOA is 1.38×10^{43} molecule^{2}cm^{-5}, where 93% and 73% of the total O_4 is distributed below the altitude of 10 km and 5 km, respectively. In particular, signals by the changes of O_4 are strong below 5 km, where aerosol transports are observed frequently. The vertical distributions of other atmospheric components are taken from the US standard atmosphere 1976 (United States Committee on Extension to the Standard Atmosphere, 1976). The vertical distribution of trace gases and aerosol in the troposphere are interpolated in the 0.1 km resolution from the sea level to 5 km.

2.2. DOAS analysis for O4I estimation

Table 1 summarizes the absorption cross sections of trace gases used as inputs for the radiance simulations and the DOAS spectral analysis. At wavelengths of 340, 360, 380, and 477 nm, the O_4 absorption cross section from Hermans *et al.* (1999) is used in this study. O_3 absorption cross sections at three different temperatures (223, 243, and 273 K) and NO_2 absorption cross sections at two different temperatures (220 and 294 K) are used to account for the amounts in the stratosphere and the troposphere. The radiance information obtained from the RTM simulation is analyzed to derive the O_4 SCDs using WinDOAS software (van Roozendael and Fayt, 2001) before O4I estimation. To analyze the simulated radiances, the spectrum calculated without all atmospheric gases and aerosol are used as the Fraunhofer reference spectrum (FRS). The simulated spectra are fitted simultaneously with the absorption cross sections of all trace gases listed in Table 1 and FRS in the respective wavelength range of 335-350, 350-370, 370-390, and 460-486 nm, using a nonlinear least squares method (Platt and Stutz, 2008).

On the other hand, the O4I from OMI standard product of cloud (OMCLDO2) (e.g., Accarreta *et al.*, 2004; Sneep *et al.*, 2008) is used to adopt the AEH retrieval for case study. The OMCLDO2 basically used the cross section database from Newnham and Ballard (1998) considering the temperature dependence by interpolating to representative atmospheric temperature of 253 K (Accarreta *et al.*, 2004). For this reason, there can be systematic difference between the O4I from OMCLDO2 and direct estimation from the observed radiance spectra in this present study. Figure 2 shows the O_4 SCD from OMCLDO2 and those directly retrieved from radiance spectrum over all observed OMI pixels on March 31, 2007 over East Asia. Similar to the DOAS analysis using the simulated spectra for a look-up table (LUT) calculation, OMI observed radiance spectra are fitted with the Ring spectrum and the FRS in addition to the absorption cross sections in Table 1 within the same wavelength window. Before the spectral fitting, the NO_2 and O_3 cross sections are I_0 corrected, and the Ring spectrum (Fish and Jones, 1995), accounting for the effects of the rotational Raman scattering due to air molecules, is calculated using the WinDOAS software (van Roozendael and Fayt, 2001). After the fitting, the noise level of residual spectrum is estimated to be on the order of 10^{-3} for the radiance spectrum at 477 nm from OMI measurements. The O_4 SCDs with the fitting error less than 1% is used for the comparison. From this figure, a systematic difference between the two different fitting results is less than 1%, although the cross section databases for fitting are different. From this result, the effect of cross section database difference is negligible when the same observation data was used. Furthermore, the DOAS analysis for LUT calculation can be used to compare the O_4 SCD from OMCLDO2.

Figure 3 shows the comparison of the O_4 SCD at 477 nm from LUT with the dimension as in Table 2 against OMCLDO2 for aerosol and cloud free pixels in year 2005. The LUT of O_4 SCD is estimated by the DOAS analysis using simulated radiance from VLIDORT with various geometries as shown in Table 2. The clear sky region is selected for the Pacific Ocean with cloud fraction less than 0.02 from OMI observation. The surface albedo is assumed to be 0.05, which is similar to the minimum Lambertian equivalent reflectance (LER) over clear ocean surface (e.g., Kleipool *et al.*, 2008). Because the standard product of the O_4 SCD is only estimated at the 477 nm band, the results can be compared only at this band. To minimize the DOAS fitting error, the observed data from OMI is selected by the fitting precision less than 2% and the quality flags for spectral fitting are also considered. As shown in Figure 3(a), the coefficient of determination (R^2) is 0.864 with a slope of 1.050, and the LUT exhibits a ratio of 0.86±0.05 to the values obtained from OMI standard values. Despite the statistically significant R^2 and slope values between the two values, there exists negative bias by about 14%.

The bias between the retrieved from LUT and the estimated from standard product values can be attributed to the differences in the O_4 cross section data and the lack of their temperature and pressure dependence as noted from the previous works by Wagner *et al.* (2009), Clemer *et al.* (2010), and Irie et al. (2015). For this reason, ground-based measurements adopted the correction factors to cross section database. However the bias effect for the cross section difference is limited as shown in Figure 2, and the correction factor for the cross section database in the previous studies cannot be adopted to the space-borne measurements. From Kleipool *et al.* (2008), the minimum LER is defined to be the 1% cumulative probability threshold, and frequent LER value are typically higher than minimum LER over clear ocean, although cloud screening was perfectly executed before LER calculation. To account for the difference between simulated and observed SCD, the LUT was re-calculated by changing condition to the surface albedo of 0.10. Although assumed surface albedo is higher than minimum LER from Kleipool *et al.* (2008), the surface albedo of 0.10 is realistic value for ocean surface albedo at mid-latitude (e.g., Payne, 1972). The corrected result is shown in Figure 3(b), with the R^2 of 0.865 similar to that before the correction, while the negative bias is removed to 0.98±0.05 and the regression line slope is 1.123. Although the comparison result is not perfect, the calculation by the VLIDORT simulates the satellite observation and can be used for sensitivity tests and case studies to retrieve aerosol height.

3. Sensitivity test

3.1. Sensitivity of the O4Is to the AEH

The sensitivity of the O4I to the AEH is investigated for its absorption bands at 340, 360, 380, and 477 nm. Figure 4 shows the O4I as a function of the AEH and the three different aerosol types of MITR, WASO and COPO at 360, 380, and 477 nm, respectively. The vertical error bar represents the fitting error estimated by the residual spectra from the DOAS fitting (e.g., Stutz and Platt, 1996). For the calculation shown in the figures, the following geometries are assumed: solar zenith angle (SZA) of 30 degrees, viewing zenith angle (VZA) of 30 degrees, and relative azimuth angle (RAA) of 100 degrees. Note that insignificant SCD value was estimated at 340 nm due to the large spectra fitting error. In these three figures, the O4Is show the AEHs ranging from 1.0 to 5.0 km for the AODs of 1.0 and 2.5 at 500 nm, which could be due to the existence of thick aerosol layers. For the sensitivity result, the decrease rate of the O4I value in the 1 km interval of AEH ($-dO_4/dZ$) is defined as equivalent O4I difference converting from O4I difference between neighbor AEH in same AOD condition.

The O4Is are estimated at 360 and 380 nm band as shown in Figure 4(a) ~ (f).

The O4I is significantly decreased with increasing AEH at 360 and 380 nm for all aerosol types. However negative O4Is are occasionally estimated at 360 nm. Furthermore the fitting errors are too large to estimate the AEH, which range from 160 to 410 at 360 nm and from 350 to 1060 at 380 nm. From large fitting error with small O4I, the fitting results are insignificant at these two absorption bands.

On the other hand, the sensitivity of the O4I at 477 nm is a significant variable to estimate AEH. The mean value of $-dO_4/dZ$ is estimated to be 87, 290, and 190 for the MITR, WASO, and COPO when the AOD is 1.0, respectively. The mean value of $-dO_4/dZ$ on the AOD of 2.5 is estimated to be 94, 362, and 213 for the MITR, WASO, and COPO, respectively. The calculated $-dO_4/dZ$ are significantly larger than the mean O4I fitting errors of 58, which implies that the O4I at 477 nm is useful in estimating the AEH. The small fitting errors at 477 nm are due to the larger O_4 absorption and less interferences by other trace gases in this spectral window.

Figure 5 shows $-dO_4/dZ$ as changing viewing geometries. As enlarging geometrical path length for viewing geometry, $-dO_4/dZ$ also increases because the path length through the aerosol layer is also increased. The mean value of $-dO_4/dZ$ including all cases of AEH is estimated to be 90 to 326 at SZA of 30.0 degree and VZA of 30.0 degree, while it is estimated to be 265 to 485 at SZA of 60.0 degree and VZA of 60.0 degree. Although aerosol scattering angle is changed by SZA and VZA, the O4I sensitivity to AEH is generally increased to increasing optical path length to the viewing geometries. From this result, the accuracy for the AEH retrieval is potentially better for large zenith angle cases than for low zenith angle cases.

3.2. Error analysis

Errors are also estimated in terms of key variables in the estimation of the O4I at 477 nm, with the variables and their dimensions as summarized in Table 3. For the error analysis of AEH retrieval, characteristics for all of extinction properties are essential to consider. In this study, errors are analyzed in terms of AOD, aerosol vertical distribution, particle size and SSA for aerosol amount and properties. Surface albedo variation is also considered to represent surface condition. To estimate the error amount, the AEH error is converted from the half of O4I difference between adding and deducting perturbation of variables as shown in equation (1).

$$\varepsilon(\mathrm{Z}) = \left| \frac{\mathrm{O4I}(\mathrm{x}+\delta\mathrm{x},\mathrm{Z}) - \mathrm{O4I}(\mathrm{x}-\delta\mathrm{x},\mathrm{Z})}{2.0 \times \mathrm{dO4I}/\mathrm{dZ}(\mathrm{x},\mathrm{Z})} \right| \quad (1)$$

where $\varepsilon(\mathrm{Z})$ is the AEH error amount due to variable of error source, x, in AEH of Z, and $\delta\mathrm{x}$ is perturbation of AEH retrieval error source. The $\varepsilon(\mathrm{Z})$ value also depends on viewing geometries. Therefore $\varepsilon(\mathrm{Z})$ is represented for specific geometries together with averaging over all geometries.

3.2.1. AOD

The O4I at 477 nm has sensitivity not only for AEH but for AOD as shown in Figure 4(g) ~ (i). Because the radiance extinction by aerosol changes depending on AOD, the optical path length of TOA radiance is also affected by AOD. For different AODs (τ_a), the O4I at AEHs of 1.0 and 3.0 km is shown in Figure 6 for the same geometry assumed in Figure 4. From OMI standard products, the expected error of the AOD over ocean is the larger of 0.1 or 30% for absorbing aerosol, and the larger of 0.1 or 20% for non-absorbing aerosol (Torres *et al.*, 1998, 2002). For this reason, the uncertainty of AOD is assumed to be 0.1 in this study, although uncertainty of AOD would be larger than the assumed value for large AOD. The decreasing rate of the O4I ($-dO_4/d\tau_a$), which defines O4I reduction with AOD increase by 0.1, is found to be larger for the AEH at 3.0 km than for that at 1.0 km. Among the three aerosol types, the $-dO_4/d\tau_a$ is found to be the least for the WASO, which has stronger scattering characteristics than other two aerosol types. In addition, the sensitivity for WASO showed negative $-dO_4/d\tau_a$ for small AOD at low AEH, which has small shielding effect with large enhancement effect, due to the large SSA of WASO. The mean $-dO_4/d\tau_a$ values are estimated to be 1.2%, 0.9%, and -0.1% for the AEH of 1.0 km as the AOD changes by 0.1 for the MITR, COPO, and WASO, respectively, whereas they are estimated to be 2.3%, 2.1%, and 1.0% for the AEH of 3.0 km with respect to the same AOD changes for the three different type, respectively.

Figure 7 shows the expected error in AEH due to retrieval uncertainty of AOD from observation. Because O_4 concentration exponentially decreases as the atmospheric altitude increases, the sensitivity to AEH becomes weak at high AEHs. In addition, aerosol signal is relatively weak for low AOD. From these reasons, the AEH retrieval error due to AOD uncertainty is maximized for the high AEH with low AOD cases for all aerosol types. The maximum retrieval error is 2.0, 0.7, and 4.4 km for COPO, WASO, and MITR for the case at AEH of 4.0 km and AOD of 0.4, which is least sensitive case for AEH. For AOD of 0.4, however, the retrieval error due to AOD uncertainty is 0.3, 0.2, and 0.4 km for COPO, WASO, and MITR for the case at AEH of 1.0 km. Except for AEH lower than 4 km and AOD larger than 0.4, the retrieval error of AEH is less than 1.0 km for all viewing geometries and all aerosol types.

Furthermore, the AEH error for AOD uncertainty is also dependent on viewing geometries. From previous studies, the error for cloud height information depends on the observation geometries due to changing average optical path length (Accarreta *et al.*, 2004; Chimot *et al.*, 2015). Moreover, the retrieval error sensitivity for observation geometries is also found in aerosol height estimation by O_2-A band (Sanders *et al.*, 2015). Similar to these previous studies, the AEH error becomes larger for short light paths and smaller for long paths. Figure 8 shows the viewing geometry dependence of AEH error for AOD of 1.0. With the increase in effective optical path length, the radiance signal from aerosol is also enhanced. In general, the AEH error decreases with increasing viewing geometries. For WASO case, however, the AEH error is smaller for short path length than long path length in low AEH case. For thin aerosol layer situation, the radiance is enhanced by scattering aerosols which results in increasing optical path length. In the small SZA and VZA, aerosol layer effectively brings enhancement effect. With increasing SZA and VZA, however, the shielding effect due to aerosol layer enhances because radiance has to pass through long path through aerosol layer. For this reason, the smallest error case is inflection point of AOD sensitivity, which corresponds to turnaround point between with larger shielding effect than enhancement effect.

3.2.2. SSA

Torres *et al.* (1998) showed that the result of the SSA from OMI can be overestimated due to the cloud contamination, although aerosol retrieval algorithm considers the existence of cloud in sub-pixel. Furthermore, the SSA varies widely for different aerosol types. Therefore, the sensitivity of O4I to the SSA variation is estimated for the same geometries used in the previous tests. To estimate O4I sensitivity to the SSA variation, the imaginary part of refractive index value corresponding to 10% variability for SSA is changed after fixing the real part of refractive index. The mean O4I changes by 106, 282, and 205 for MITR, WASO, and COPO, respectively, with respect to its SSA deviation by 10%. To compare the difference for WASO and COPO, it is proportional to the absolute values of the SSA for all simulated cases. In addition, the difference for MITR is smaller than those for COPO, because less fraction of back scattering in coarse mode particle makes less sensitive to O4I.

Figure 9 shows the AEH error due to the SSA variation by 10%. Because of the low sensitivity characteristics of AEH as shown in section 3.1, thus large errors are shown for high AEH and low AOD cases. However, the AEH errors are less than 1 km for COPO aerosol type. For AOD of 1.0, the AEH error due to SSA variation is estimated to be 610 to 900 m for the COPO type. Furthermore, the error is calculated to range from 270 to 1220 m and from 930 to 1400 m for COPO and WASO type, respectively, if AEH is 3 km, frequently assumed reference altitude in aerosol retrieval algorithm (e.g., Torres *et al.*, 1998). For MITR, dust-like type aerosol, AEH error, which ranges from 410 to 1430 m for AOD of 1.0, is generally the largest compared to those of other aerosol types. In general, uncertainty of aerosol optical properties is large for thin aerosol layer case, thus that of the AEH is as well.

3.2.3. Particle size

Aerosol particle size has noticeable effects on the phase function, thus the directional scattered intensity. However, most of aerosol retrieval algorithm assumes aerosol particle size depending on its type as an input parameter to RTM calculation. Although aerosol type is categorized, however, physical properties of aerosol can be changed according to the source type and transport characteristics. In the OMI aerosol algorithm, size distribution is one of error sources for the AOD (Torres *et al.*, 2002).

Figure 10 shows the AEH error due to particle size change. For error estimation, mode radius difference for number size distribution is assumed to be ±20%, which corresponds to larger range by 4 times than those from the error budget study for OMI standard product (Torres *et al.*, 2002). Overall, O4I difference is within the order of 100. The coarse mode aerosol, MITR in this study, results in the largest O4I difference for all cases, thus the largest AEH error for MITR which is estimated to range from 0.2 to 2.7 km. On the other hand, the error ranges from 0.03 to 0.5 km and from 0.2 to 1.9 km for WASO and COPO, respectively. The largest AEH errors for the three aerosol types are estimated for the case with AOD of 0.4 and AEH at 5.0 km.

3.2.4. Surface Albedo

As the surface albedo affects the $-dO_4/dZ$, the sensitivity of the O4I is also tested with respect to the surface albedo difference of 0.02. The difference of climatological surface albedo between that obtained from the total ozone monitoring spectrometer (TOMS) and the global ozone monitoring experiment (GOME) was known to be up to 0.02 (Koelemeijer *et al.*, 2003). Table 4 shows the sensitivity of the O4I with respect to the change in the surface albedo. The absolute difference of O4I due to surface albedo variation is below 85. Because aerosol layer attenuates the reflected radiance from surface, the absolute difference of O4I value decreases as aerosol amount increase. Furthermore, it is found that the difference of O4I due to surface albedo change is larger for the non-absorbing aerosol than the absorbing aerosol, because absorbing aerosol attenuates the reflected radiance more than the non-absorbing aerosol. In terms of AEH change, the O4I difference increases as AEH increase. For low AEH case, optical path length of reflected radiance from surface to aerosol layer is relatively short as compared to high AEH case. For this reason, the O4I sensitivity for surface albedo is reduced by high concentration of aerosol near the surface for the low AEH case.

Figure 11 shows the expected retrieval error of AEH due to surface albedo difference as changing AEH with respect to AOD and its types. As mentioned in previous section, the -dO_4/dZ is small in high AEH and low AOD cases. Furthermore, the albedo sensitivity increases as AEH increases and AOD decreases. As a result, the AEH error is frequently larger than 1 km for high AEH with small AOD, especially when AOD is less than 0.4. Because reflected radiance from surface is dominant for thin aerosol case, the AEH error in high AEH with low AOD shows the largest value as compared to previous error analysis. However, the AEH error sharply decreases as AOD increases and AEH decreases, when aerosol signal becomes dominant. Especially for MITR, four simulation cases, when AOD = 0.4 with AEH > 3.0 km and AOD = 1.0 with AEH = 5.0 km, show the AEH error larger than 1 km. Because -dO_4/dZ is too small in these cases, AEH retrievals in the four simulation cases show limitation as a reliable result. For COPO and WASO, however, all the cases in AEH < 3.0 km, which directly influence surface concentration, show error lower than 750 m even for the assumed AOD of 0.4. In addition, errors less than 500 m are found for AOD > 1.0 with AEH < 3.0 km.

3.2.5. Vertical distribution

Aerosol vertical distribution varies largely by distance from source, atmospheric dynamics during aerosol transport, and sink mechanism in reality. To estimate the AEH error due to variation of aerosol vertical distribution, the half-width of GDF distribution was doubled for comparison. Although it is not possible here to consider all kinds of aerosol vertical distributions due to its large variability in profile, aerosol vertical distribution by changing the half-width of GDF distribution can reflect large-scale changes in its vertical profile.

Table 5 shows the mean AEH errors between the two vertical profiles of aerosol as AOD changes. As the aerosol vertical profile is changed with increasing its widths, the difference of O4I ranges from 100 to 430. Because aerosol vertical profile simultaneously affects aerosol concentration and layer thickness, the O4I difference shows large value as the vertical distribution changes. For this reason, the AEH error is larger than 2.5 km for all aerosol types with AOD of 0.4. The estimated errors caused solely by the change between the two aerosol vertical profiles, range 1477±602, 722±190, and 671±265 m for the MITR, COPO, and WASO, respectively, for AOD greater than 1.0.

3.3. Error budget

Table 6 shows the summary of the total error budget for the AEH estimation with a list of the major error sources and their values, assuming errors in each variable in OMI standard products. To convert the O4I difference to the AEH error, the difference of O4I due to the respective error source is divided by that from the change of the AEH in each bin of the AOD and AEH as shown in section 3.2, with the simulation cases over 58,800 runs listed in Table 3 to calculate mean and standard deviation of errors. Because of weak signal sensitivity to AEH for AOD of 0.4 and AEH at 5.0 km as shown in the previous section, this simulation case is omitted in calculating statistical values for error budget. In summary, the total number of aerosol simulations for the combination of AOD and AEH includes 39 cases.

The mean errors from 10% variation in the SSA for all of the variable conditions in Table 3 correspond to 726, 576, and 1047 m for the MITR, COPO, and WASO, respectively. For the total error budget calculations, however, SSA change by 5% was used according to Torres *et al.* (2007), which reported the variation of the SSA less than 0.03 for the given aerosol type. The error from the vertical distribution is estimated to be 720, 1480, and 690 m for the COPO, MITR and WASO, respectively.

The errors from SSA and aerosol profile shape are the two important error sources in estimating the AEH, followed by the errors related to AOD and surface albedo. From these results, the errors of the AEH due to the error from OMI AOD of 0.1 and the surface albedo of 0.02 are less than 300 m for WASO and COPO, and about 400 m for MITR. However, the AEH error from surface albedo is important for cases with low AOD at high AEH, which is surface reflectance dominant case.

The mean errors from 20% variation in the aerosol particle size are 726, 576, and 1047 m for the MITR, COPO, and WASO, respectively. Torres *et al.* (2002) assumed the variation of size distribution to be 5%. Thus, for the total error budget calculations assuming 5% change in the particle size, the AEH errors are less than 100 m. In addition, the errors in the O4I, and thereby the AEH, are associated with the variations in the column amounts and the differences in the absorption cross section of each fitted trace gas for the spectral analysis. The variations in the column amounts of trace gases and the differences in the absorption cross section values do not affect significantly in calculating the O4I. However, the O_4 vertical column density is changed by the change in atmospheric pressure. In East Asia, the surface pressure over ocean is 1010.9±29.6

(3-sigma) hPa from NCEP Reanalysis 2 data since 2004. In clear case, the difference of O4I due to the ±3% for pressure variation is 3.4±0.1% in all geometries.

Furthermore, the AEH error in terms of inaccurate spectral wavelength calibration is estimated based on the assumed errors of ±0.02 nm, which corresponds to 0.1 pixels for OMI. Although it is well known that the accuracy in the spectral wavelength calibration before the DOAS fitting affects the trace gas SCD retrieval, the errors in the O4I associated with the wavelength shift of the sub-pixel scale are estimated to be negligible due to the broad O_4 absorption band width around 477 nm.

Finally, the total error budget in the AEH retrieval is estimated based on the error analysis with respect to error sources. Note that the result of error analysis explains about 50% for SSA and 25% for size parameter in calculating the total error budget. Overall, the total error budget in the AEH retrieval is estimated to be 739, 1276, and 846 m for the COPO, MITR, and WASO, respectively, with the exception of the contribution of the errors in the aerosol vertical profiles. Therefore accurate assumption for optical properties of aerosol is essential to develop the retrieval algorithm of aerosol height.

4. Case study

To demonstrate the feasibility from real measurements, the AEHs are derived using hyperspectral data from OMI. OMI channels are composed of UV-1 (270-314 nm), UV-2 (306-380 nm), and a visible wavelength range (365-500 nm) with a spectral resolution (FWHM) of 0.63, 0.42, and 0.63 nm, respectively (Levelt *et al.*, 2006). The spatial resolution is 13 km × 24 km at nadir in "Global Mode". In the present study, the OMI spectral data over the visible wavelength range are used to derive the O4I at 477 nm and the AEH information.

Figure 12 shows an AEH retrieval algorithm for the case study. In retrieving AEH, AOD is obtained from MODIS standard product (e.g., Levy *et al.*, 2007). Although OMI aerosol product provides AOD at 500 nm, AOD from OMI was partially affected by aerosol height and suffered from cloud contamination due to its large footprint (Torres *et al.*, 2002). For this reason, AOD from MODIS was allocated to the OMI pixels as a reference AOD for the AEH retrieval. For type selection, the AE from MODIS and AI from OMI is respectively used for the information of size and absorptivity, to classify aerosol type into four following the method from Kim *et al.* (2007) and Lee *et al.* (2007). After determining AOD and aerosol type, LUT, which is generated as functions of geometries (SZA, VZA, and RAA), aerosol types and AODs, is used to determine the AEH information by comparing simulated with the measured O4I value. The variables and their dimensions for the LUT calculations are listed in Table 7. Due to the limitation of the accuracy of aerosol type classification and those of AOD over land, this study estimates the AEH only over ocean surface. Although temporal and spatial variation of surface albedo influences the AEH result from error study, surface albedo is assumed to be a fixed value of 0.10, which is used in the sensitivity study. Even if the surface albedo is changed but known, the qualitative conclusion here is not affected. For case study, the LUT of O4I is developed by the aerosol model based on AERONET data over East Asia. Extensive AERONET dataset over East Asia are used to provide represent aerosol optical properties for the LUT calculation.

Figure 13 shows the results of the retrieved AEH during the Asian dust event on March, 31, 2007. MODIS products of AOD and FMF on this date show that thick dust layer with the AOD up to 1.0 from China to the Yellow Sea [Figure 13(b)] and the FMF ranging from 0.2 to 0.4, indicating the dominance of coarse-mode particles [Figure 13(c)]. Using the basis of the current algorithm with the pre-determined AOD and type, the mean retrieved AEH is 2.3±1.3 km over 647 pixels in East Asia [Figure 13(d)]. The retrieved result is compared with the backscattering intensity from the CALIOP observation over the Yellow Sea as shown in Figure 13(e). From CALIOP observation, the aerosol layer height over Yellow Sea is located at around 1 km altitude for most of observed regions. Over the Yellow Sea domain in 35~40 ˚N and 120~130 ˚E, the AEH from OMI is 1.5±1.1 km over 166 pixels, which is within 1 km difference from the CALIOP. From the retrieved result, the retrieved AEH is successfully retrieved within expected error, and the current algorithm quantitatively estimates the AEH over East Asia.

Figure 14 is another case study of the retrieved AEH on February, 21, 2008. MODIS products of AOD and FMF on this date show thick anthropogenic aerosol transported with the AOD ranging from 0.6 to 1.0 [Figure 14(b)] and the FMF ranging from 0.8 to 1.0 [Figure 14(c)] all over the Yellow Sea. The mean retrieved AEH is 1.4±1.2 km over 1480 pixels in East Asia as shown in Figure 14(d). On this date, CALIOP passed over coastal line between China and Yellow Sea. The aerosol layer height ranged from 0.5 to 2.5 km during the overpass over East Asia as shown in Figure 14(e). The AEH from OMI is 0.6±0.4 km over 601 pixels in 30~40 ˚N and 120~125 ˚E. Contrary to large spatial variation of the AEH from CALIOP, the AEH from OMI shows spatially stable values on this date.

Figure 15 shows the scatter plot of AEH between CALIOP and OMI on the dates in Table 8, which lists aerosol transport cases over East Asia with simultaneous observations by OMI and CALIOP in 2007 and 2008. The AEH from CALIOP is estimated by the data from vertical profile of aerosol extinction coefficient at 532 nm. Because the O4I sensitivity for AEH is not large at AEH higher than 4 km, the comparison test was limited to cases with AEH less than 4.5 km from OMI. For data collocation, the latitude and longitude difference between two sensors are within 0.25 degree. Figure 15(a) shows the comparison of AEH from OMI and CALIOP with MODIS AOD larger than 0.5. It is assumed that the reference expected error (EE) is 1 km (Fishman *et al.*, 2012). Almost 60% of retrieved pixel shows the AEH result within the EE. Because of large AEH error for low AOD, the accuracy of AEH result from OMI is poor. Furthermore, this case study assumes constant surface albedo value over ocean. However, ocean surface albedo is also changed by turbidity due to sediments and ocean surface due to wind. For this reason, the AEH error is exaggerated for low AOD cases. If lower of AOD for the comparison is set to be 1.0, the proportion of pixel within EE improves up to 80% as shown in Figure 15(b). Furthermore, the correlation of the AEH between the two sensors is improved with the regression line slope of 0.62 and the correlation coefficient (R) of 0.65 for thick aerosol layer cases. Therefore, the AEH algorithm from OMI provides the reasonable information about the parameter of aerosol vertical distribution, if accurate aerosol model is provided for the forward calculation.

5. Summary & discussion

The sensitivities of the O4I at 340, 360, 380, and 477 nm bands are investigated with RTM calculations to derive the AEH using the space-borne hyperspectral data. Among these O_4 absorption bands, the O4I at 477 nm is considered to be suitable for the AEH retrieval. In addition to the AEH, AOD, aerosol type, aerosol vertical profile, particle size, and surface albedo are also found to have effects on the O4I at 477 nm, while the spectral calibration and cross section of the atmospheric gases have negligible effects on the O4I. The major error source for the AEH retrieval is found to be the uncertainty in SSA, which leads to the AEH error ranging from 270 to 1400 m with the SSA perturbation by 10%. In addition, the profile shape is also a major error source for the AEH estimation. According to the error estimations, the total errors are 739, 1276 m, and 846 m for absorbing, dust, and non-absorbing aerosol, respectively, due to combined uncertainties of the variation from AOD, SSA, particle size, and surface albedo.

In addition to the sensitivity analysis, an algorithm for the AEH derivation is developed for the first time based on a LUT that consists of the O4I in terms of the AEH, AOD, aerosol types, surface albedo, and measurement geometries. After the determination of AOD and aerosol types from the MODIS, the AEH value is derived over East Asia by the current algorithm using OMI measurement data. Considering the accuracy of AOD and aerosol types, the result is shown over ocean surface. From several cases for the long-range transport of aerosol over East Asia, the derived AEH shows reasonable value as compared to aerosol layer height from CALIOP with the correlation coefficient of 0.62 for AOD larger than 1.0. In addition, 80% of estimated AEH from OMI showed error less than 1 km in AEH.

There are many works to be done to improve the newly introduced algorithm as it requires the products from MODIS to determine the AOD and aerosol types prior to the AEH retrieval. The vertical distribution and the optical properties of the aerosol need to be quantified using the combination of observation database, such as MPLNET and AERONET. Furthermore, the spatial variation of the AOD, surface pressure and the contamination by the cloud in the sub-pixel scale need to be investigated as they are also thought to affect the retrieved results. If the surface reflectance can be characterized with sufficient accuracy, the retrieval of the AEH can be extended to over land. In addition, the O4I method in this study can be applied to the surface pressure estimation in clear regions.

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Tables

Table 1. The database of cross section used for DOAS fitting analysis.

Species	Temperature (K)	Reference
O_3	223, 243, and 273	Bogumil *et al.* (2001)
NO_2	220 and 294	Vandaele *et al.* (1998)
O_4	298	Hermans *et al.* (1999)

Table 2. Dimensions of LUT for the clear sky comparison.

Variable name	No. of Entries	Entries
SZA	7	0, 10, 20, 30, 40, 50, 60 degrees
VZA	7	0, 10, 20, 30, 40, 50, 60 degrees
RAA	10	0, 20, 40, 60, 80, 100, 120, 140, 160, 180 degrees

SZA : Solar zenith angle, VZA : Viewing zenith angle, RAA: Relative azimuth angle.

Table 3. Dimensions of simulation cases for the error analysis of the AEH retrieval.

Variable name	No. of Entries	Entries
SZA	7	0, 10, 20, 30, 40, 50, 60 degrees
VZA	7	0, 10, 20, 30, 40, 50, 60 degrees
RAA	10	0, 20, 40, 60, 80, 100, 120, 140, 160, 180 degrees
AOD	5	0.4, 1.0, 1.6, 2.5, 3.0
AEH	8	1.0, 1.2, 1.6, 2.0, 2.4, 3.0, 4.0, 5.0 km
Aerosol Model	3	MITR, WASO, COPO
Surface Albedo	1	0.10

AOD: Aerosol Optical Depth, AEH : Aerosol Effective Height

Table 4. Absolute difference of O4I for changing surface albedo by 0.02.

	MITR	WASO	COPO
Maximum	81	85	76
Case [AOD,AEH]	[0.4,5.0]	[0.4,5.0]	[0.4,5.0]
Minimum	8	11	1
Case [AOD,AEH]	[3.0,1.0]	[3.0,1.0]	[3.0,1.0]
Mean±Standard Deviation	38±22	37±20	20±21

Table 5. The error for AEH due to the change in aerosol vertical distribution.

Reference shape (half-width=1 km)	MITR (half-width=2 km)	WASO (half-width=2 km)	COPO (half-width=2 km)
Error for AEH [m]	1477±602	671±265	722±190

Table 6. Summary of error sources and total error budget for the AEH retrieval.

<table>
<tr><th>Error source</th><th>MITR</th><th>WASO</th><th>COPO</th></tr>
<tr><td>AOD
(ΔAOD = 0.2)</td><td>387±740 m</td><td>105±131 m</td><td>218±358 m</td></tr>
<tr><td>SSA
(10% change)</td><td>726±537 m</td><td>1047±194 m*</td><td>576±332 m</td></tr>
<tr><td>Surface Albedo
(Δα = 0.02)</td><td>438±762 m</td><td>199±241 m</td><td>154±274 m</td></tr>
<tr><td>Particle Size
(20% change)</td><td>352±174 m</td><td>72±56 m</td><td>315±213 m</td></tr>
<tr><td>Atmospheric Gases</td><td colspan="3">< 5 m</td></tr>
<tr><td>Atmospheric Pressure**
(ΔP = 3%)</td><td colspan="3">3.4±0.1% (O_4 SCD)</td></tr>
<tr><td>Instrument
(Shift : 0.02 nm)</td><td colspan="3"><10 m</td></tr>
<tr><td>Total Error</td><td>1276 m</td><td>846 m</td><td>739 m</td></tr>
</table>

* Calculation results for the SSA decrease by 10%.

** For clear sky calculation.

Table 7. Dimensions of LUT for the AEH algorithm using OMI.

Variable name	No. of Entries	Entries
SZA	7	0, 10, 20, 30, 40, 50, 60 degrees
VZA	7	0, 10, 20, 30, 40, 50, 60 degrees
RAA	10	0, 20, 40, 60, 80, 100, 120, 140, 160, 180 degrees
AOD	13	0.0, 0.2, 0.4, 0.6, 0.8, 1.0, 1.3, 1.6, 1.9, 2.2, 2.5, 3.0, 5.0
AEH	16	0.0, 1.0, 1.2, 1.4, 1.6, 1.8, 2.0, 2.2, 2.4, 2.6, 2.8, 3.0, 3.5, 4.0, 5.0, 10.0 km
Aerosol Model	3	Dust, Carbonaceous, Non-absorbing [Climatology over East Asia site of AERONET]

Table 8. List of aerosol transport cases and its period for comparison.

Case	Period
1	March, 28, 2007 - April, 2, 2007
2	May, 5, 2007 - May, 10, 2007
3	May, 25, 2007 - May, 26, 2007
4	February, 19, 2008 - February, 21, 2008
5	April, 3, 2008 - April, 5, 2008
6	May, 28, 2008 - May, 31, 2008
7	December, 4, 2008 - December, 7, 2008

Figures

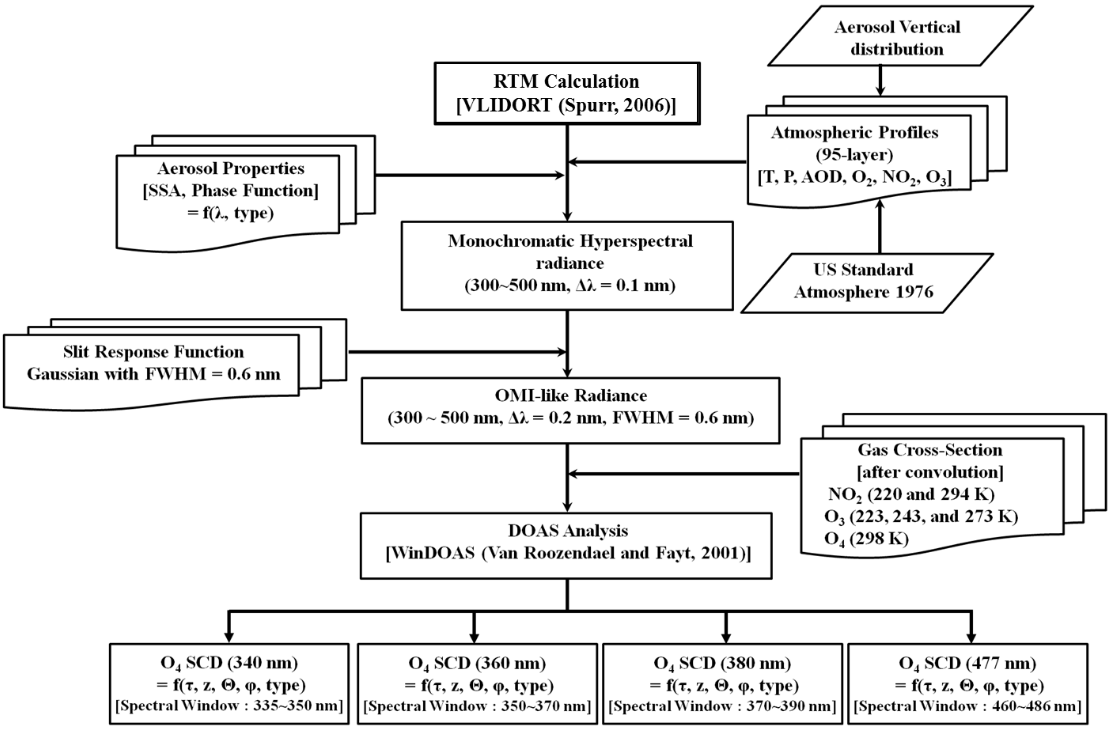


Figure 1. Flowchart of the simulated O_4 SCD estimation.

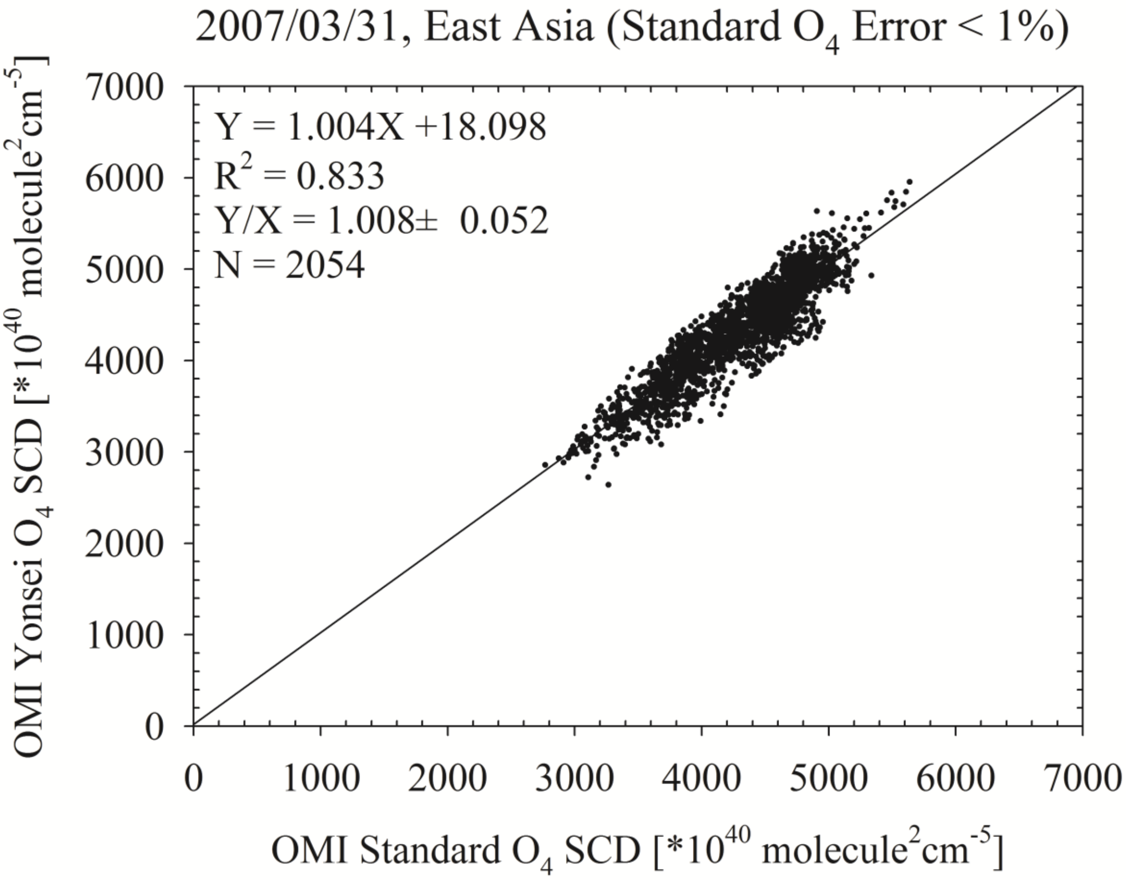
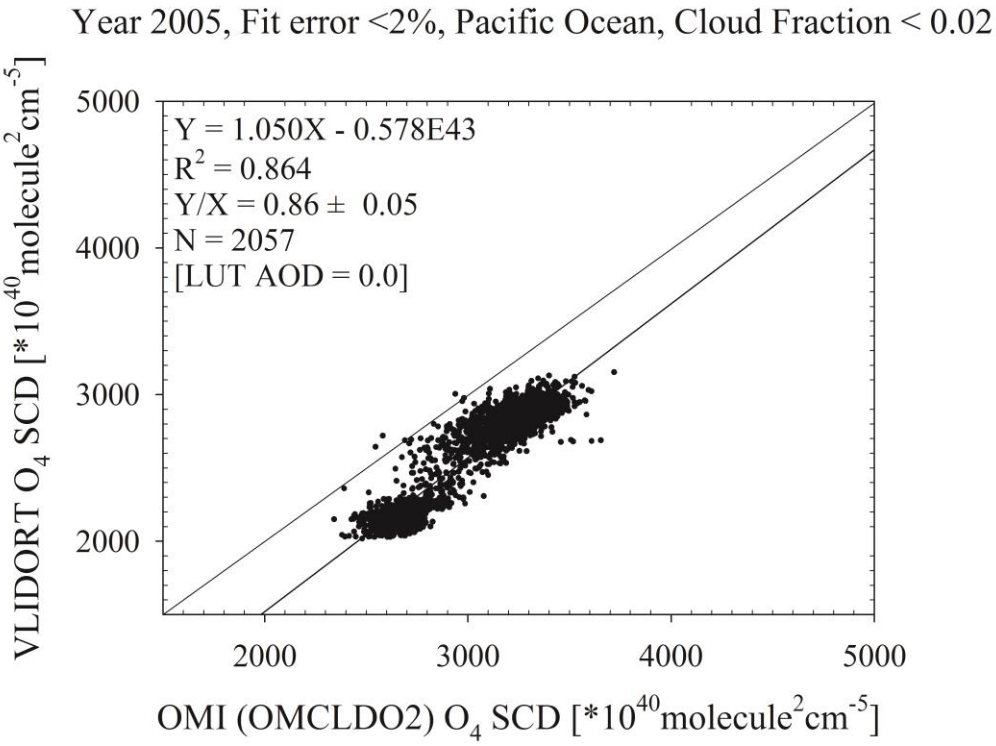


Figure 2. Comparison of O_4 SCD directly retrieved from OMI radiance with the OMI standard product on March 31, 2007.

(a)



(b)

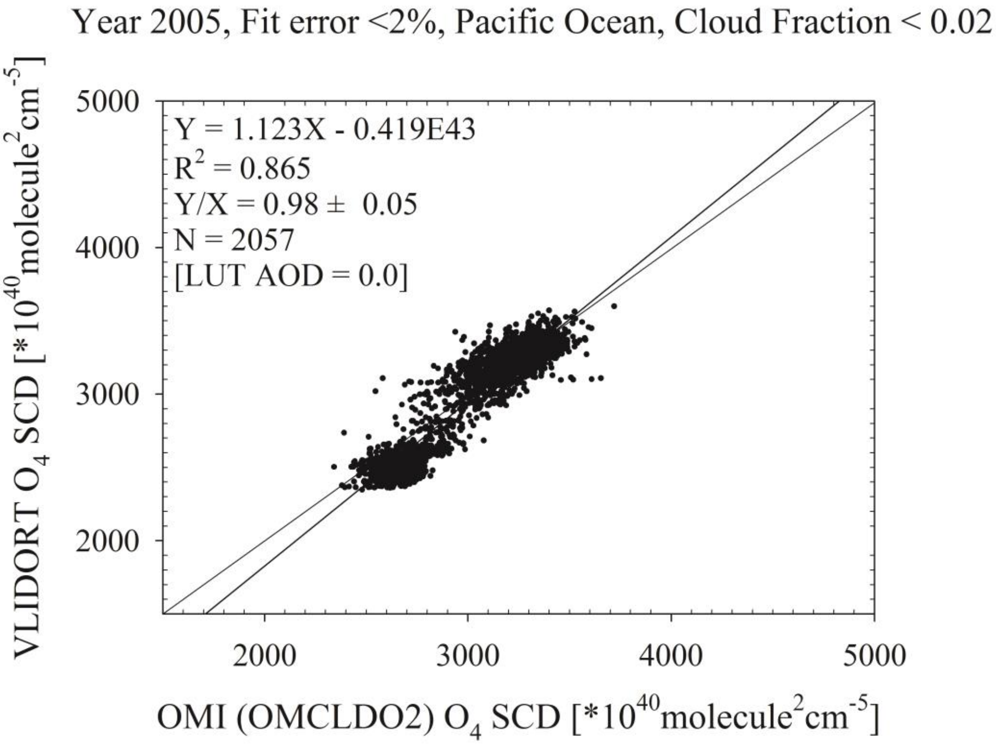


Figure 3. Comparison of the O_4 SCD at 477 nm between the OMI standard product and the calculated value from LUT (a) before and (b) after correction of LER.

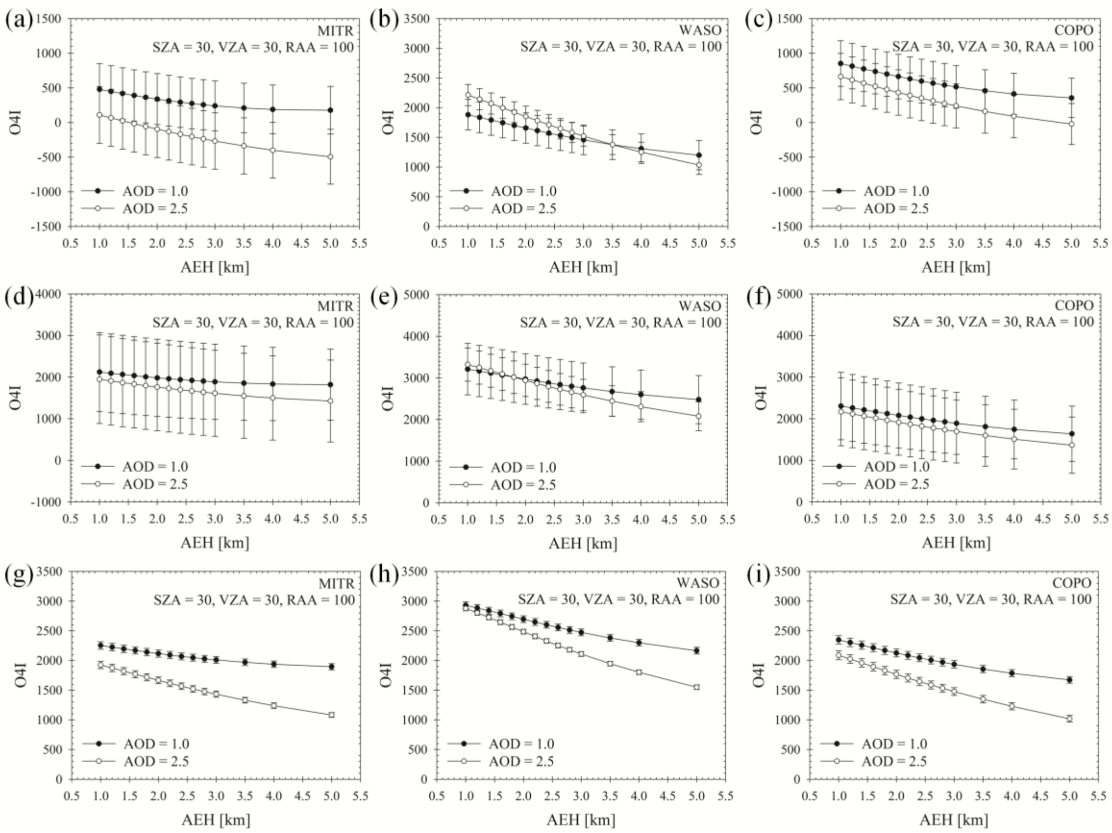


Figure 4. The O4I at 360 nm band for (a) MITR, (b) WASO, and (c) COPO, (d) at 380 nm band for MITR, (e) WASO, and (f) COPO, and (g) at 477 nm band for MITR, (h) WASO, and (i) COPO as a function of AEH.

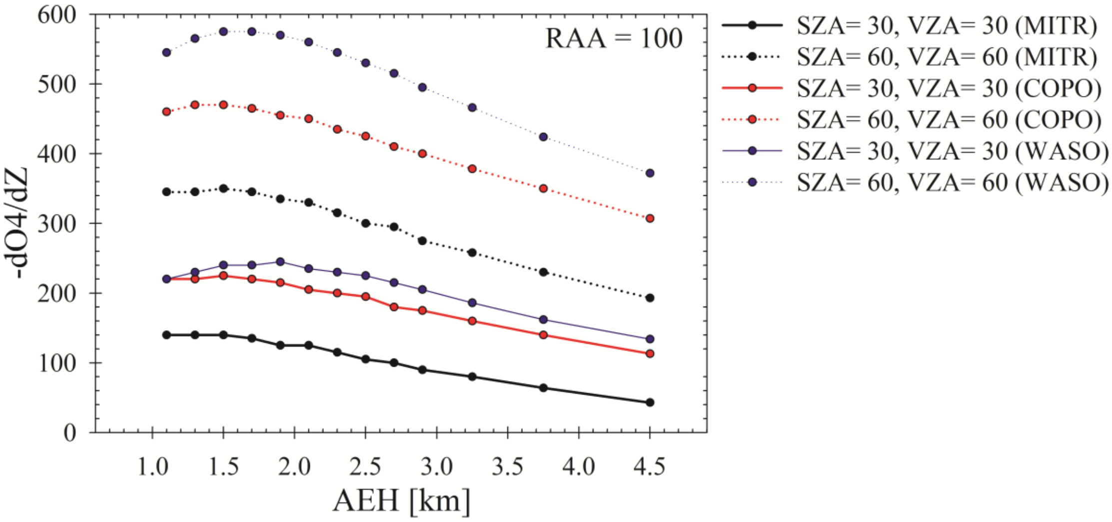
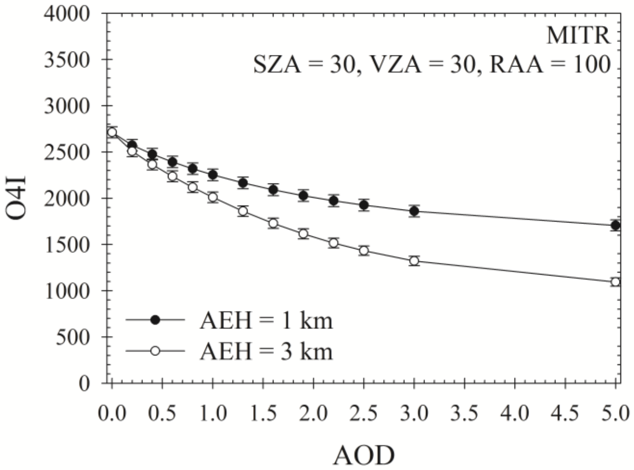
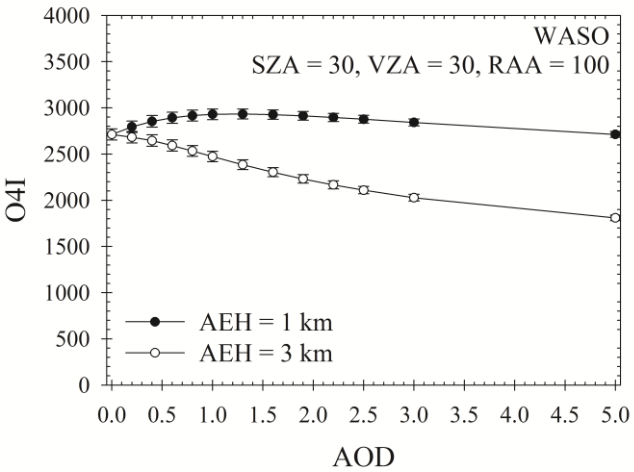


Figure 5. The AEH sensitivity to O4I (-dO4/dZ) with changing observation geometries at 477 nm.

(a)



(b)



(c)

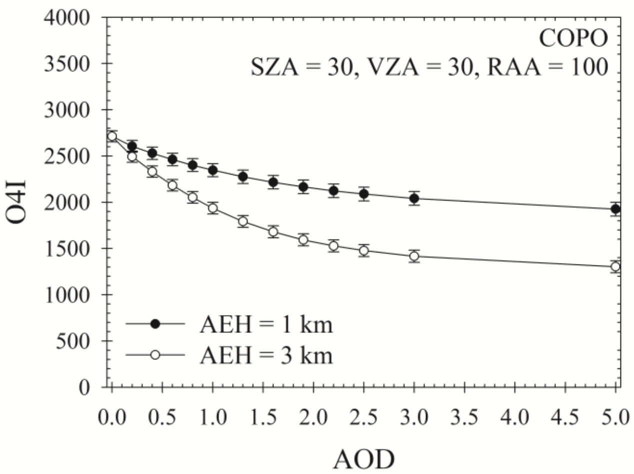
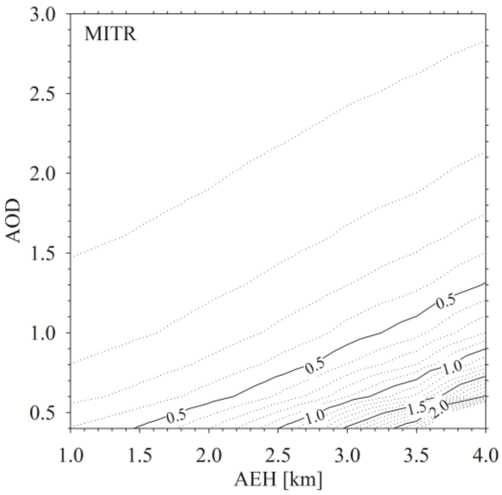
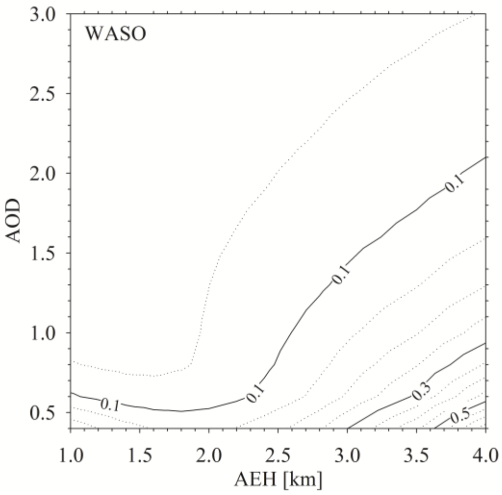


Figure 6. The O4I of (a) MITR, (b) WASO, and (c) COPO types as a function of AOD.

(a)



(b)



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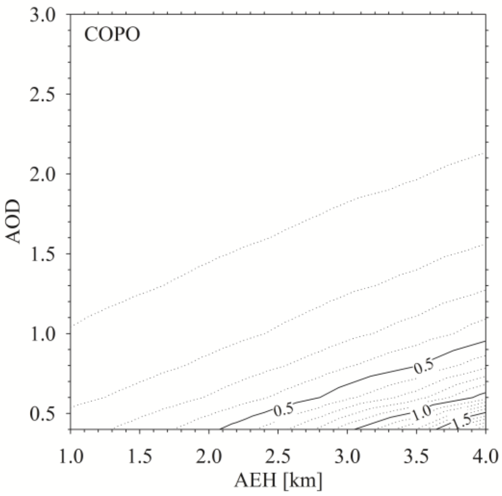
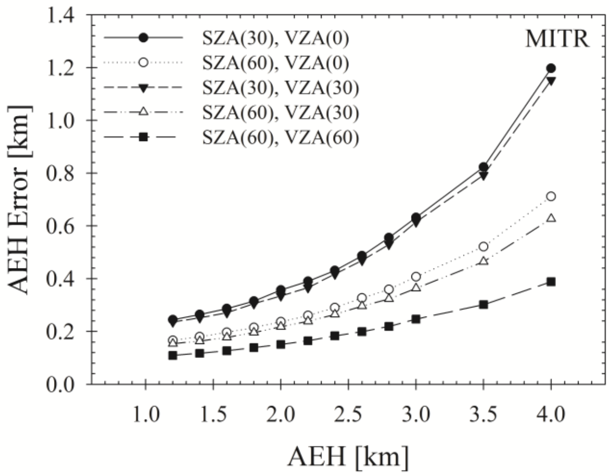
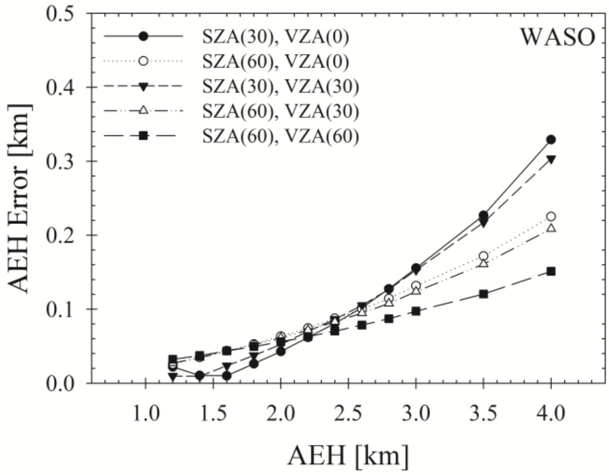


Figure 7. AEH error of (a) MITR, (b) WASO, and (c) COPO for the AOD difference of 0.1 as a function of reference AEH and AOD.

(a)



(b)



(c)

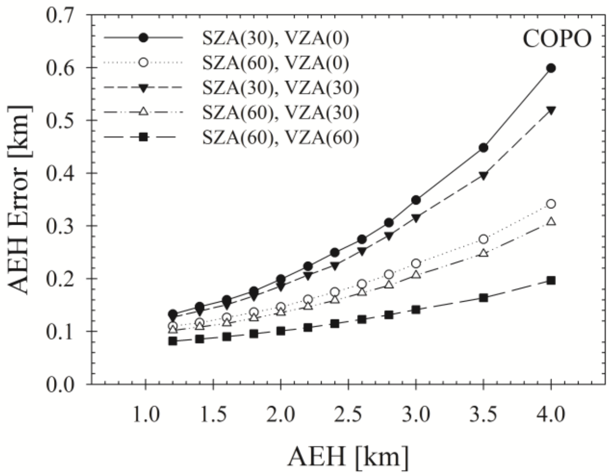
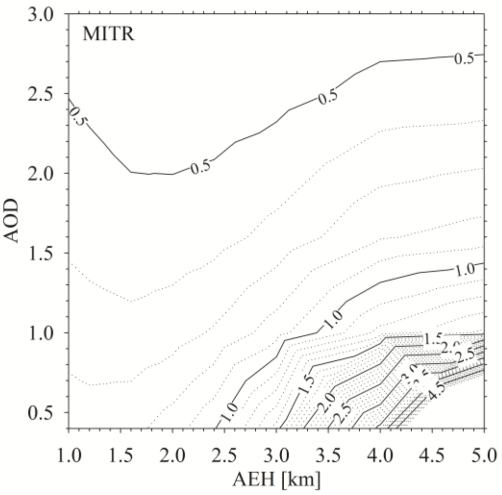
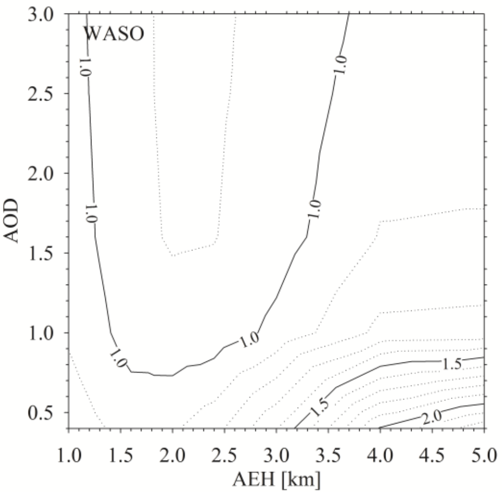


Figure 8. AEH error of (a) MITR, (b) WASO, and (c) COPO for the AOD difference of 0.1 as changing viewing geometries.

(a)



(b)



(c)

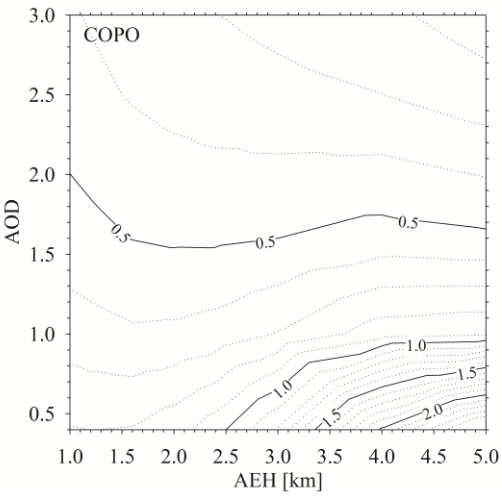
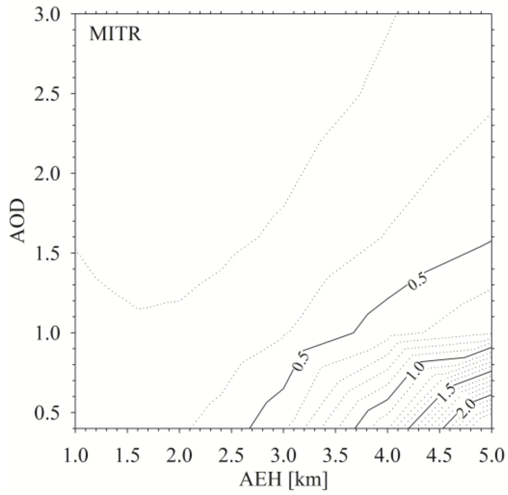
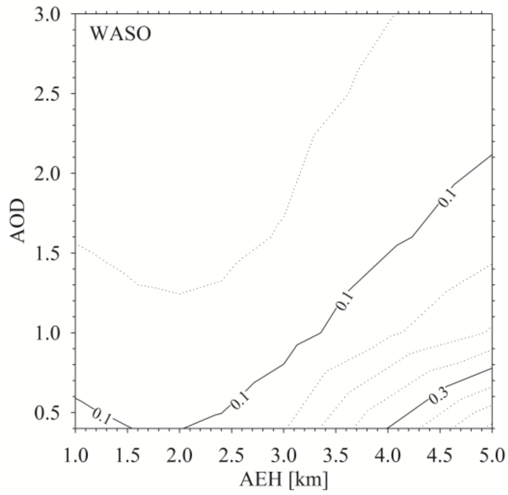


Figure 9. Same as Figure 7 except for SSA difference of 10%.

(a)



(b)



(c)

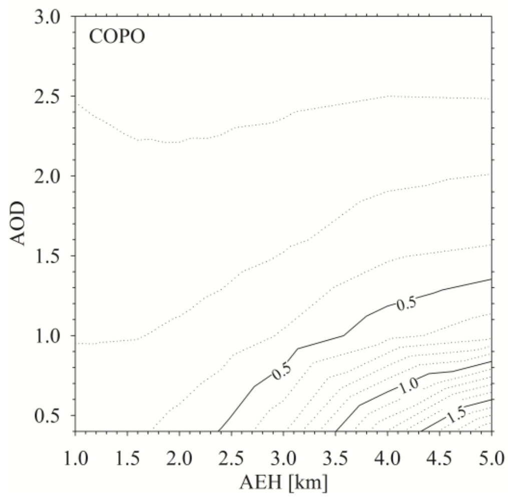
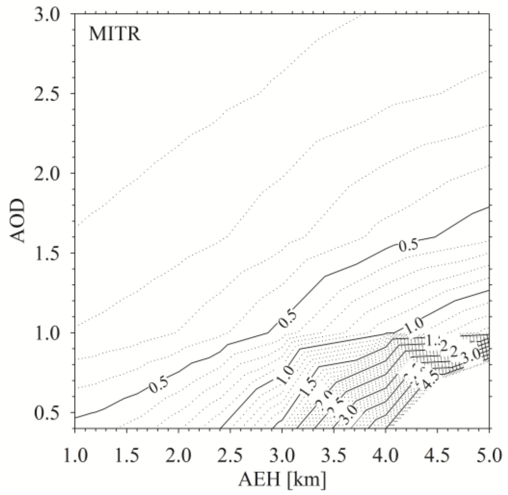
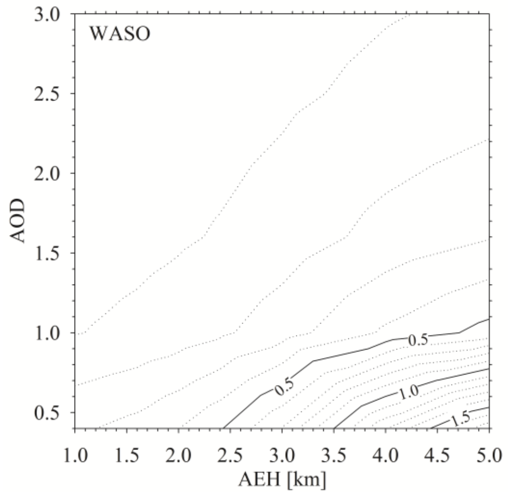


Figure 10. Same as Figure 7 except for particle size difference of 20%.

(a)



(b)



(c)

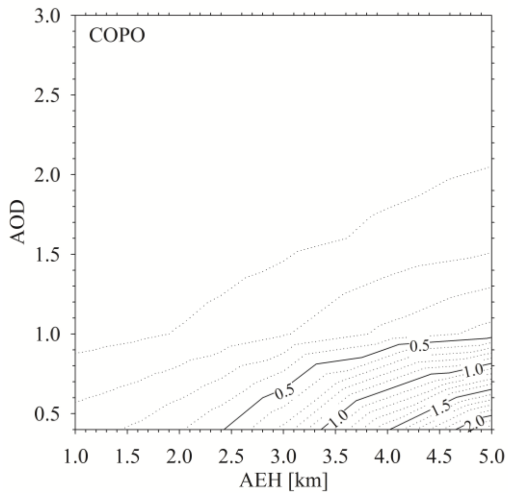


Figure 11. Same as Figure 7 except for surface albedo difference of 0.02.

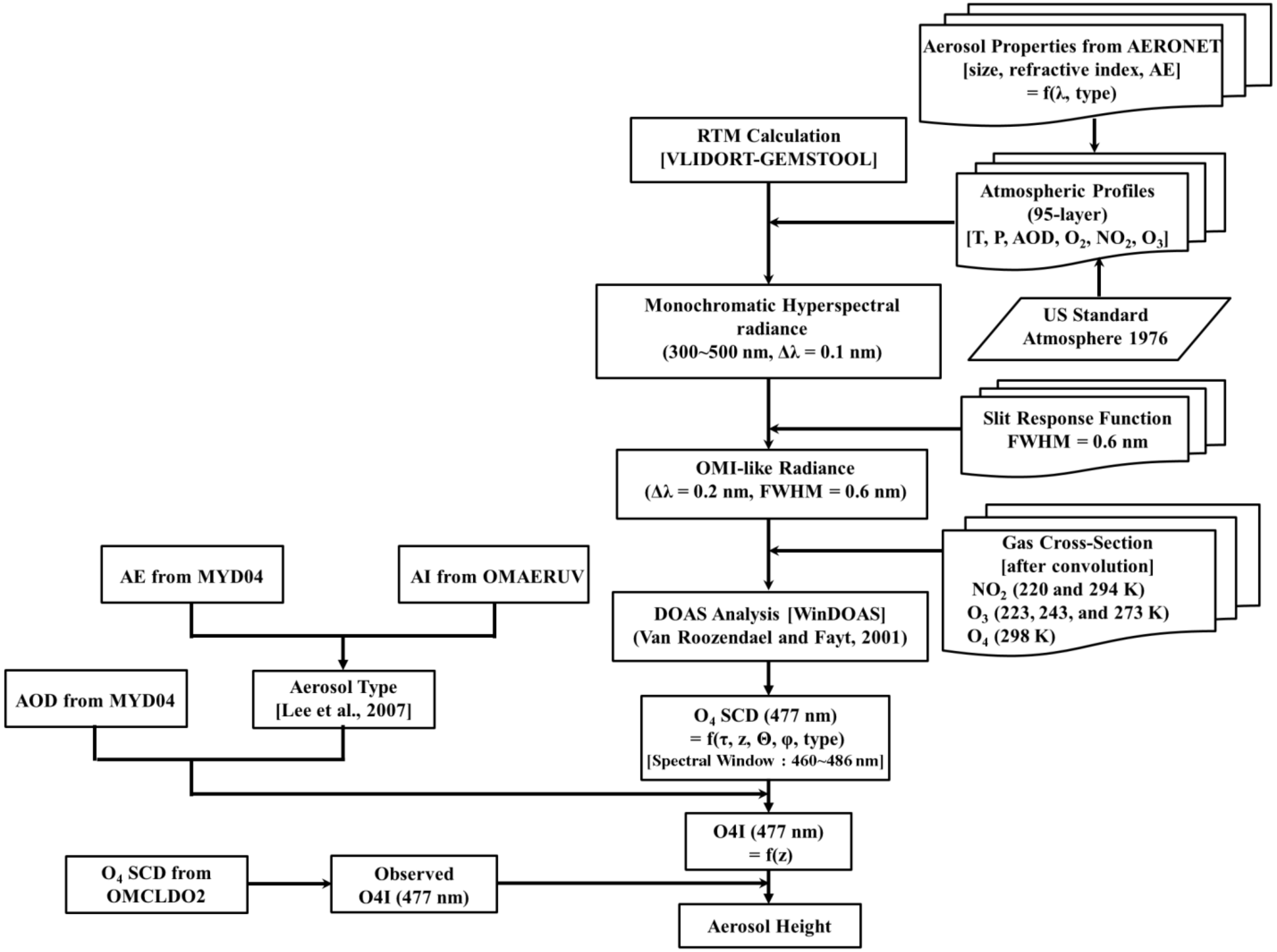


Figure 12. Flowchart of the retrieval algorithm for AEH from OMI observation.

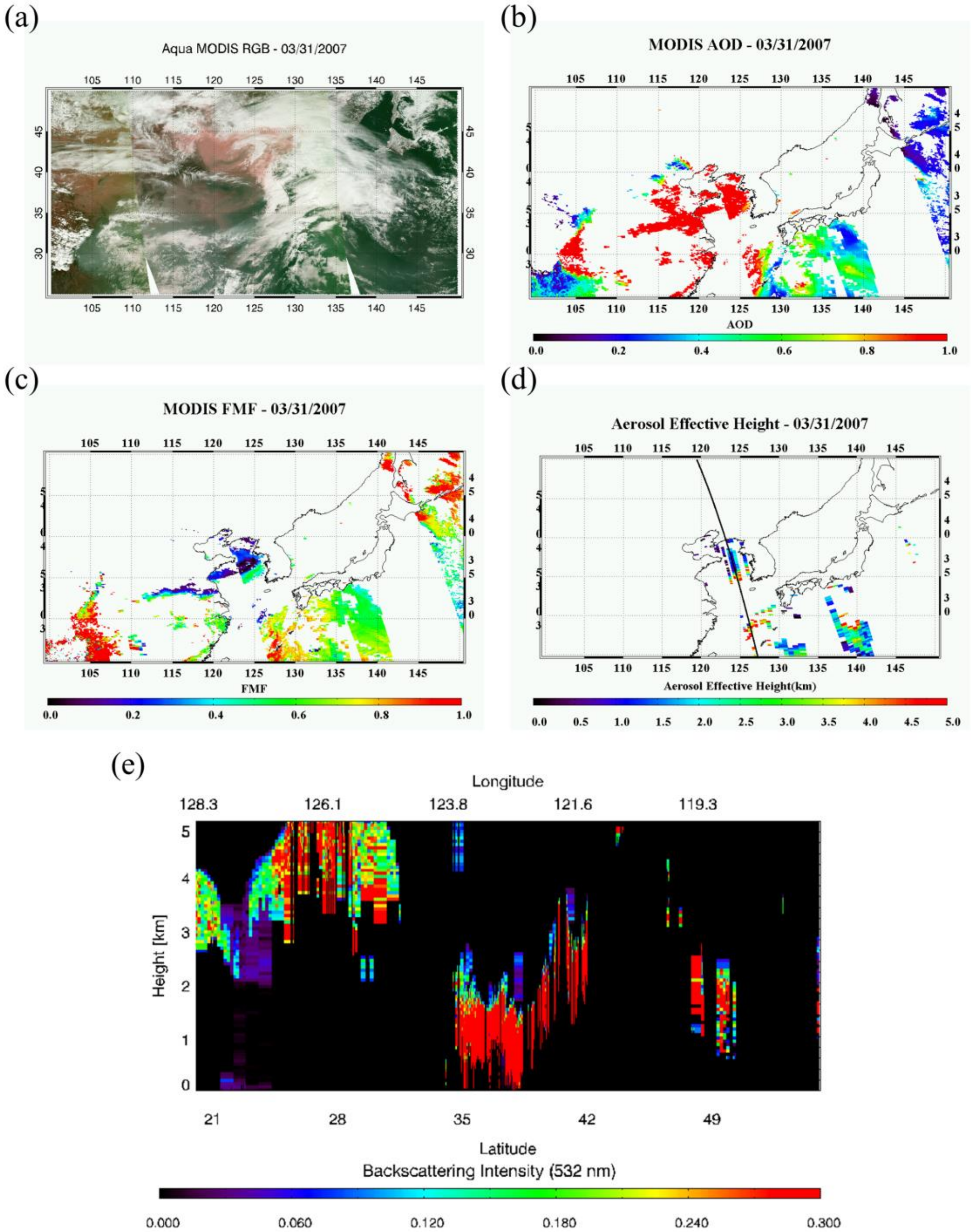


Figure 13. (a) MODIS RGB, (b) AOD, (c) FMF, and (d) AEH distribution from OMI over East Asia, and (e) Backscattering Intensity at 532 nm from CALIOP observation over Yellow Sea on March 31, 2007.

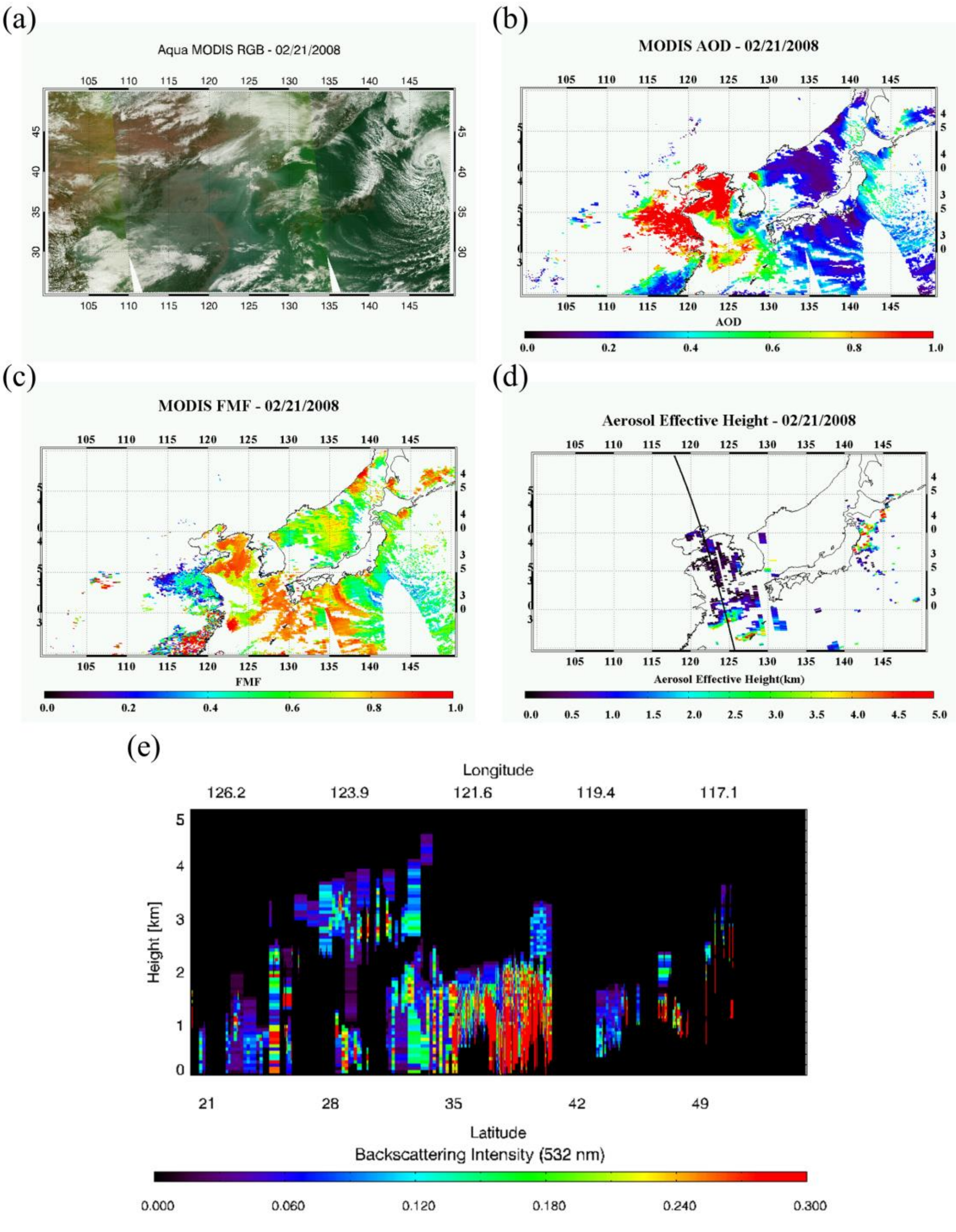
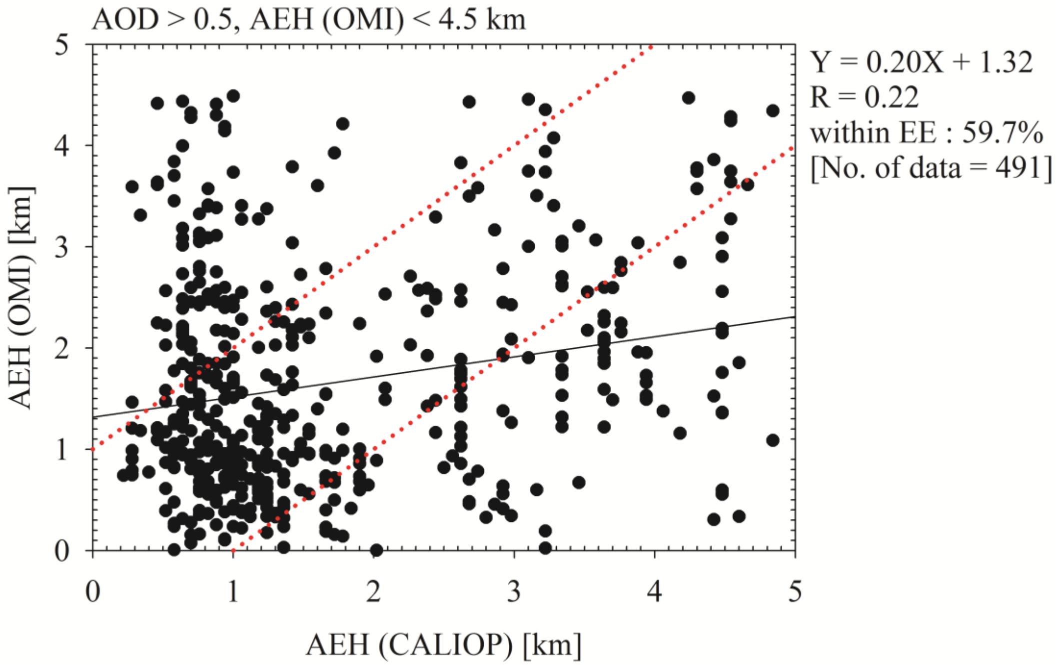


Figure 14. (a) MODIS RGB, (b) AOD, (c) FMF, and (d) AEH distribution from OMI over East Asia, and (e) Backscattering Intensity at 532 nm from CALIOP observation over coastal region of China on February 21, 2008.

(a)



(b)

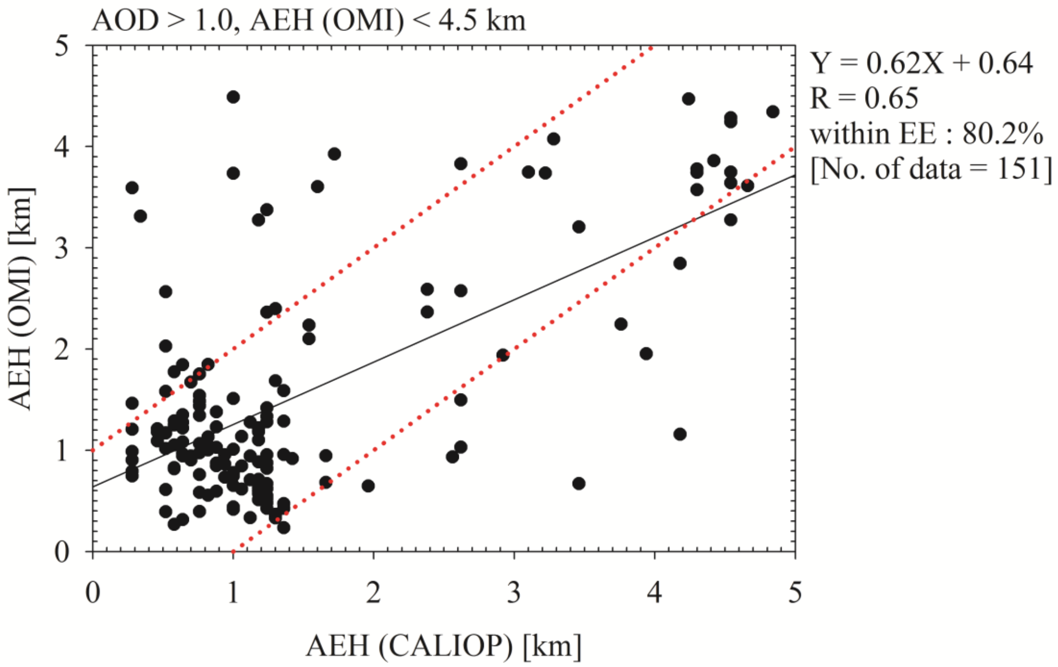


Figure 15. Comparison of AEH from OMI with CALIOP with (a) AOD > 0.5, and (b) AOD > 1.0 for aerosol transport cases in 2007 and 2008 over East Asia.