Use of 315 nm Channel Data of the Sky Radiometer to Estimate the Columnar Ozone Concentration: A Preliminary Study

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Abstract

The sky radiometer is a ground-based instrument implemented to investigate the characteristics of not only aerosols but also clouds and water vapor. This study is the first attempt to bring this instrument in use to estimate the columnar ozone concentration (U). A method and preliminary results related to the estimation of U using the 315-nm-channel data of the sky radiometer are presented. The proposed method consists of calculating the calibration constant for direct intensity at 315 nm wavelength, F_0 (315 nm), using in situ observation data, which is an alternative of the traditional Langley method, and estimating U . The temporal values of U at Chiba, Japan for the period of January–April, 2013 were estimated, and they were compared with values observed by the ozone monitoring instrument (OMI). The agreement was satisfactory during the initial observation period; however, the values from the sky radiometer were observed to be gradually overestimated with time. The study suggests that the temporal change of F_0 (315 nm) is the important factor to be considered while estimating U values for long-term observations and that F_0 (315 nm) must be periodically determined. The study further discusses the uncertainties associated with the estimated U due to the uncertainties in the values of input parameters.

Keywords sky radiometer; ozone; calibration constant

1. Introduction

Ozone is one of the most important constituents of our atmosphere, although it comprises less than 0.001 % of air by volume. Ozone in the lower atmosphere is an air pollutant with effects on the degradation of air quality; however, the stratospheric ozone prevents harmful ultraviolet-B (UV-B) radiation (280–320 nm) from reaching the Earth's surface (Taalas et al. 1997). In addition, it impacts tropospheric and stratospheric chemical processes (Levy 1971), as well as global

climate change (Shindell et al. 2006). Therefore, ozone variations are of worldwide concern in recent decadal years, and several instruments are being deployed for its observation.

For the last several years, ozone has been observed from space using instruments measuring backscattered radiances, such as Total Ozone Mapping Spectrometer (TOMS), Ozone Mapping Instrument (OMI), Solar Backscatter UltraViolet (SBUV), and Global Ozone Monitoring Experiment (GOME) (<http://mascuv.gsfc.nasa.gov>). The Superconducting Submillimeter-Wave Limb-Emission Sounder (SMILES), which was successfully launched and attached to the Japanese Experiment Module (JEM) on the International Space Station (ISS) on 25 September 2009 (Kikuchi et al. 2010), also observes ozone from

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space. Moreover, it is being regularly observed using ground-based instruments, such as the Dobson spectrometer and the Brewer spectrophotometer. Those instruments observe either the direct sun irradiance or the zenith sky radiance at a few discrete UV wavelengths (Fioletov et al. 1999). Furthermore, some other ground-based instruments measuring the spectral UV irradiances are in use for observing ozone. For example, Gao et al. (2001) used direct irradiances of double pair wavelengths (305 and 325 nm and 311 and 332 nm) observed by the ultraviolet multi-filter rotating shadow-band radiometer (UV-MFRSR) to retrieve the columnar ozone concentration (U). In addition, some studies (e.g., Stames et al. 1991; Masserot et al. 2002; Houet and Brogniez 2004) have shown that U can be retrieved using global irradiances in the UV spectral range.

Since ozone absorbs radiation within the UV spectral range, one may directly use the relation between U and the observed direct intensity within this spectral range to retrieve U using the well-known Beer–Lambert law. Such technique can be expected to be reliable, and its accuracy can be easily evaluated. Moreover, it does not require a complicated radiative transfer simulation like the studies that use global irradiances to retrieve U . However, for the instrument that observes direct irradiance at the surface in terms of current or voltage, one needs to know the calibration constant for direct irradiance before using this technique. The most common method to determine such calibration constants is the Langley plot, which needs a very stable atmospheric condition. Therefore, such calibration is generally performed on high mountains, which is expensive, laborious, and inconvenient.

In this study, we propose a method to determine the calibration constant for direct irradiance at 315 nm wavelength F_0 (315 nm), by avoiding laborious and inconvenient calibration techniques, and a method to determine U using the sky radiometer observations. As the sky radiometer data have been successfully used to retrieve aerosol optical parameters (Nakajima et al. 1996), cloud optical parameters (Kikuchi et al. 2006), and precipitable water content (Campanelli et al. 2010), our procedure that is described in this paper may help the sky radiometer community gather information on an additional atmospheric parameter, i.e., U , using the same instrument in future. There are other important implications of this study: though the data of U can be obtained from space-based instruments, the retrieval of U from the sky radiometer can reach a higher temporal resolution, thus allowing the study of the diurnal variation. At the same time, such

data from the sky radiometer can also be used to validate the products of numerical simulations, as well as space-based observations.

2. Instrumentation and data

The two models of the sky radiometer (POM-01 and POM-02), manufactured by PREDE Co., Ltd., Japan, are being operated around the world. The improved model (POM-02) is equipped with annexed interference filters for the selection of working wavelengths for up to 11 channels. The standard wavelengths for the sky radiometers operated by the SKYNET network (<http://atmos.cr.chiba-u.ac.jp/>) are 315 nm, 340 nm, 380 nm, 400 nm, 500 nm, 675 nm, 870 nm, 940 nm, 1020 nm, 1627 nm, and 2200 nm. Among them, measurements at wavelengths ranging from 340 to 1020 nm, except 940 nm, are specifically used to study aerosols. Similarly, the 315 and 940 nm wavelengths are useful to study ozone and precipitable water content, respectively. The remaining two wavelengths of 1627 and 2200 nm can be employed to study the optical characteristics of clouds, as well as aerosols. At present, the channels with wavelengths from 340 to 1020 nm, except 940 nm, are used as standard channels to retrieve aerosol optical parameters under the SKYNET network. The POM-02 instrument has two detectors with different fields of view for wavelengths less than or equal to 1020 nm and wavelengths greater than 1020 nm. It is worth mentioning that data of the 315 nm channel were less accurate to estimate U because of the unsatisfied optical transmittance in the past. To improve this issue, the manufacturer has developed and revised a part of the optical system, and the latest version of the POM-02 instrument can make it possible. This study is the first attempt to use the data of such instruments to estimate U .

One such POM-02 instrument belonging to Chiba University (hereafter referred as Chiba_POM-02) was operated at the Meteorological Research Institute (MRI), Tsukuba from 13 December, 2012 to 8 January, 2013 as a part of a periodic calibration using a standard instrument of MRI. During this period, the Chiba_POM-02 was run in “sun photometer” mode to observe direct intensities every minute at the wavelengths described above. One of the POM-02 instruments belonging to MRI (hereafter referred as MRI_POM-02) has been simultaneously operated in “normal mode” to observe direct intensities every 10 min, as well as sky radiances at each wavelength stated above.

As discussed in Section 3.1, we need direct inten-

sity, aerosol optical thickness, ozone optical thickness, and Rayleigh scattering optical thickness at 315 nm to estimate F_0 (315 nm), which is required in turn to estimate U . As further discussed in Section 3.1, the aerosol optical thickness at 315 nm is obtained from a Mie calculation (Bohren and Huffman 1983) using the refractive index at the nearest wavelength, i.e., 340 nm, and the volume size distribution. To derive such volume size distribution and the refractive index using the sky radiometer data, the data of sky radiances observed at different scattering angles are essential (Nakajima et al. 1996). Since the information of only direct intensities observed by the Chiba_POM02 were not enough to derive the volume size distribution and the refractive index, such volume size distribution and refractive index were derived from the data of collocated MRI_POM02. Note that the MRI_POM02 observed direct intensities, as well as sky radiances, because it was operated in “normal mode.” Input of such information in the SKYRAD.pack software (Nakajima et al. 1996) allows one to obtain the volume size distribution and spectral refractive indices, as well as the spectral optical parameters of aerosols. Hence, we also had to use the MRI_POM02 data. Note that if Chiba_POM-02 had been run in “normal mode” to observe direct intensities, as well as sky radiances, the data of MRI_POM-02 would not have been necessary. Therefore, we want to emphasize that the use of two sky radiometers is not a mandatory condition for the purpose of determining F_0 (315 nm).

Furthermore, we used U data directly observed by the Dobson spectroradiometer at the Aerological Observatory close to the MRI from 13 December, 2012 to 8 January, 2013. Data of this period, i.e., data collected at Tsukuba, were used to derive F_0 (315 nm) with the procedure described in Section 3.1. After 8 January, 2013, the Chiba_POM-02 instrument was moved to Chiba University and was operated in “normal mode” until the middle of May, 2013. In Chiba University, data were collected at 3 min time intervals. Data observed at Chiba University were used to estimate U values using the procedure described in Section 3.2. Along with the ground-based observation data, the U values observed by the ozone measuring instrument (OMI), which is onboard the Aura satellite, were also used to compare with the results obtained from the sky radiometer. The details of the OMI sensor and its products can be found in the respective publications (e.g., Bhartia et al. 2002). This study uses OMI level 3 products with spatial resolution of $0.25^\circ \times 0.25^\circ$.

3. Method

The estimation of U using the sky radiometer data comprises two parts: the determination of F_0 (315 nm) and the determination of U .

3.1. Determination of F_0 (315 nm)

F_0 (315 nm) can be determined from the traditional Langley method. But this technique can be expensive and laborious since it needs a very stable atmospheric condition. As an alternative of this traditional method, this study proposes a technique of implementing an improved Langley (IL) method (Tanaka et al. 1986) to determine F_0 (315 nm). Note that the IL method is common for the standard wavelengths of aerosol observation of sky radiometer (Nakajima et al. 1996; Campanelli et al. 2004), for which the observation of sky radiances is possible along with direct irradiances.

The method of determining F_0 (315 nm) uses three kinds of in situ observation data: direct irradiance at 315 nm, direct irradiances and sky radiances of the standard wavelengths for aerosol observation, and U from any standard ground-based instrument, such as the Dobson spectrometer or the Brewer spectrophotometer. The observed direct irradiances and sky radiances at standard wavelengths for aerosol observation can be implemented in the SKYRAD.pack software to determine the aerosol size distribution, the spectral refractive indices, and other aerosol optical parameters. The obtained aerosol size distribution and refractive index at the wavelength nearest to 315 nm (340 nm in this study) can be used to determine the aerosol optical thickness (τ_{aerosol}) at 315 nm using a Mie code. We calculated τ_{aerosol} (315 nm) using the refractive index at 340 nm and the aerosol size distribution, which were obtained from the SKYRAD.pack software (Ver. 4.2). A small difference in the refractive index between the wavelengths of 315 nm and 340 nm can be expected to have a negligible effect on the calculated τ_{aerosol} (315 nm) because τ_{aerosol} is a strong function of aerosol size distribution rather than the refractive index (King et al. 1978; Khatri and Ishizaka 2007). Similarly, using the known value of the atmospheric pressure corresponding the observation site, one can calculate the Rayleigh optical thickness (τ_{Rayleigh}) at 315 nm (Frohlich and Shaw 1980); the relative optical air masses for ozone (m_{ozone}) and air (m_{air}), as well as the sun–earth distance (R) can be calculated using the observation time, latitude, and longitude. The relative optical air mass, which is defined as the ratio of the optical path along the oblique trajectory to the vertical path in the zenith

direction (Iqbal 1983), mainly depends on the solar zenith angle, although its value for a specific constituent can also be appreciably influenced by the vertical distribution characteristics of that constituent (Penn-dorf 1957). In this study, m_{air} was approximated as $1 / \cos(\theta_0)$, where θ_0 is the solar zenith angle. For m_{ozone} , the expression of Robinson (1966) as given by Iqbal (1983) has been adopted.

Under the availability of the calculated and observed values of specific parameters, one can use the following relation to estimate F_0 (315 nm):

$$F = \frac{1}{R^2} F_0 \exp[-(\tau_{aerosol} m_{air} + \tau_{Rayleigh} m_{air} + K_{ozone} U m_{ozone})], \quad (1)$$

where F and K_{ozone} are the observed direct intensity and the absorption coefficient of ozone at 315 nm, respectively. For simplicity, the notation of wavelength is excluded in the above equation. In the above equation, all parameters, except F_0 and K_{ozone} , are a priori known either through calculation or observation. One may use the literature value of K_{ozone} to correlate $(\tau_{aerosol} m_{air} + \tau_{Rayleigh} m_{air} + K_{ozone} U m_{ozone})$ and $\ln(FR^2)$, while estimating F_0 . Taking into account that the value of K_{ozone} varies depending on the vertical profiles of temperature and ozone, as well as the slit function of the instrument (e.g., Kerr 2002; Bernhard et al. 2005), we simultaneously estimated K_{ozone} and F_0 rather than considering the literature value of K_{ozone} . In other words, we chose the value of K_{ozone} in such a way that the correlation between $(\tau_{aerosol} m_{air} + \tau_{Rayleigh} m_{air} + K_{ozone} \times U m_{ozone})$ and $\ln(FR^2)$ is maximum and the slope of the regression line is approximately 1.0. Figure 1 shows the scatter plot between $\ln(FR^2)$ and $(\tau_{aerosol} m_{air} + \tau_{Rayleigh} m_{air} + K_{ozone} U m_{ozone})$ for the most plausible value of K_{ozone} . Figure 1 shows that the value of K_{ozone} as 1.11 atm-cm^{-1} can produce a very strong correlation coefficient ($R^2 = 0.992$) with the slope value of 1.0. Note that K_{ozone} determined from the laboratory observed ozone absorption cross sections has been used in literature. The standard ground-based ozone observing instruments, such as Brewer and Dobson, use ozone absorption cross sections of Bass and Paur (1985) data bank. Although the actual value of K_{ozone} depends on the vertical profiles of temperature and ozone, as well as the slit function of the instrument, we compared our value of K_{ozone} with some literature values. The value of K_{ozone} at the wavelength of 315 nm at the temperature of -45°C obtained by smoothing the absorption spectra of Bass and Paur (1985) to the resolution of Brewer instrument,

Fig. 1. Scatter plot between $\ln(FR^2)$ and $(\tau_{aerosol} m_{air} + \tau_{Rayleigh} m_{air} + K_{ozone} U m_{ozone})$ for the most plausible value of K_{ozone} that should produce the strongest correlation and the slope value of the regression line as close as possible to 1.0.

which is shown by Kerr (2002) in Fig. 8, is close to our retrieved value. On the other hand, K_{ozone} at the same wavelength and same temperature determined using the same data of ozone absorption cross section is different in the study of Komhyr et al. (1993), as shown in Fig. 1 of Bernhard et al. (2005). This can be due to the different standard conditions while calculating K_{ozone} .

3.2. Determination of U

Under the condition of known F_0 (315 nm), U can be estimated from the operational data in the following ways: (i) calculate $\tau_{aerosol}$ (315 nm) from the aerosol size distribution and the refractive index of 340 nm in a similar manner as described in Section 3.1, (ii) calculate $\tau_{Rayleigh}$ (315 nm) from the information of the atmospheric pressure of the observation site using an analytical formula of Frohlich and Shaw (1980), (iii) calculate m_{air} , m_{ozone} , and R using observation time and geographical location, and finally (iv) calculate U from the following equation:

$$U = \frac{1}{m_{ozone} K_{ozone}} \left[\ln \left(\frac{1}{R^2} \frac{F_0}{F} \right) - (\tau_{aerosol} + \tau_{Rayleigh}) m_{air} \right]. \quad (2)$$

Fig. 2. (a) Comparison of columnar ozone concentrations (U) obtained from sky radiometer and OMI observations and (b) the differences between OMI and sky radiometer observed values. To compare with a single value per day of OMI observation, the sky radiometer results falling within 13:00–14:00 JST are averaged.

4. Comparison of U values obtained from the sky radiometer and OMI

Figure 2a shows the comparison of U values obtained from the sky radiometer and the OMI sensor of Aura satellite. Since OMI observations are available as a single value per day and the Aura satellite overpass time is approximately 13:38 JST, we compared the OMI observed values by taking the average of sky radiometer results within 13:00–14:00 JST. The variation in values within 13:00–14:00 JST is denoted by the standard deviation in Fig. 2a. The numbers of data points falling within this period were different for different days depending on the number of successful retrievals of aerosol optical parameters from the standard data analysis procedure of the SKYRAD.pack software. Depending on the day, the number of data points varied from 1 to 13. Note that the monthly values of F_0 of the standard aerosol observational channels were determined through the IL method and were implemented in the SKYRAD.pack software while estimating aerosol

parameters, including aerosol size distribution and refractive indices. On the other hand, the constant value of F_0 (315 nm) determined from the observation data at MRI was used for all months because the information regarding the temporal change of F_0 (315 nm) cannot be extracted from the available data sets at present. Figure 2a shows a very good agreement between the sky radiometer and OMI results in January. On the other hand, the agreement between those two independent observation systems gradually deteriorates with the Julian day. The worst scenario can be seen for the results of April, 2013. The root mean square error (RMSE) values for the months of January, February, March, and April are 0.0146, 0.0252, 0.0486, and 0.1314 atm-cm, respectively.

As shown by Fig. 2b, the difference in U values obtained from the sky radiometer and OMI observations gradually increased with the Julian day. Literature review suggests that the comparison of OMI and ground-based instruments observed values tends to show a marginal seasonal difference even though it remains unclear whether this dependence originates

Table 1. Monthly values of F_0 determined from Improved Langley (IL) method at different wavelengths along with the percentage decrement with respect to the values of January. The Julian days corresponding to each month are given in the parentheses.

Month/ Wavelength	340 nm ($\times 10^{-6}$)	380 nm ($\times 10^{-5}$)	400 nm ($\times 10^{-4}$)	500 nm ($\times 10^{-4}$)	675 nm ($\times 10^{-4}$)	870 nm ($\times 10^{-4}$)	1020 nm ($\times 10^{-4}$)
January (1–31)	7.893 0.00 %	3.064 0.00 %	1.438 0.00 %	3.315 0.00 %	4.072 0.00 %	2.706 0.00 %	1.925 0.00 %
February (32–59)	7.729 (–2.12 %)	3.01 (–1.79 %)	1.414 (–1.70 %)	3.292 (–0.70 %)	3.989 (–2.08 %)	2.694 (–0.44 %)	1.91 (–0.785 %)
March (60–90)	7.613 (–3.68 %)	2.995 (–2.30 %)	1.406 (–2.27 %)	3.28 (–1.07 %)	3.964 (–2.72 %)	2.686 (–0.74 %)	1.92 (–0.26 %)
April (91–120)	7.573 (–4.22 %)	2.966 (–3.30 %)	1.392 (–3.30 %)	3.247 (–2.09 %)	3.948 (–3.14 %)	2.668 (–1.42 %)	1.92 (–0.26 %)

from the spaceborne or ground-based observations (Balis et al. 2007). Anton et al. (2009) also showed a seasonal dependence for the comparison between OMI and ground-based observations with an excellent correlation in the winter and a fair correlation in the summer. Such seasonal dependence phenomenon might have partially contributed on the observed differences in this study, which is hard to confirm because of the limited observation period. We speculate that the possible reason for the gradual increase of the difference can be associated with the deposition of pollutant materials on the glass affecting the observed direct intensity at 315 nm or the filter degradation of the 315-nm-wavelength channel. Note that the monthly change of F_0 (315 nm) was not considered while estimating the U value; however, the monthly change of F_0 values for the standard aerosol observational channels were considered while estimating aerosol size distribution, refractive index, and other parameters. To show how F_0 can change with time, we show the monthly values of F_0 along with the percentage decrement with respect to the values of January for the standard aerosol observational channels (Table 1). Table 1 shows that the rates of the decrement of F_0 values are different for different wavelengths. In addition, we noted higher decrement rates for shorter wavelengths. A similar decrease for F_0 (315 nm) with month is possible. One can expect the closeness of U values obtained from the sky radiometer and OMI if similar decrement rate of F_0 (315 nm) is taken into account in Eq. (2). However, this information is not available at present. Nonetheless, the temporal change in F_0 (315 nm) can be the possible reason for the differences observed in Fig. 2b.

For instance, taking OMI values as reference, we roughly estimated the monthly variations of F_0 (315 nm) that might have resulted either from the deposition of pollutant materials on the glass or from filter degradation using the following equation:

$$F_{0, OMI} = F_0 e^{K_{Ozone}(U_{Ozone, OMI} - U_{Ozone, skyrad})} \quad (3)$$

Note that F_0 refers to the value given by the regression line of Fig. 1. In Fig. 3, we show F_0 (315 nm) given by Fig. 1 as “original” and those estimated using Eq. (3). While using Eq. (3), the single OMI observation per day was assumed to be constant within 13:00–14:00 JST. The average values and the standard deviations are shown in the figure along with deviation percentages from the original value. As shown in Fig. 3, F_0 (315 nm) of January deviates from the original F_0 (315 nm) by $\sim 2\%$. This difference in the F_0 (315 nm) value may resemble the differences in the U values observed by the Dobson spectrometer and OMI sensor. The percentages indicated in Fig. 3 suggest that the temporal variation of F_0 (315 nm) needs to be considered to correctly estimate U using the sky radiometer data. Therefore, the calculation of F_0 (315 nm) using the procedure of Section 3.1 at regular intervals (e.g., 6 months) is recommended to extract the information regarding the temporal change of F_0 (315 nm) while analyzing data of long-term observation. Note that such a technique is adopted by AERONET to derive F_0 values for intermediate observation days within the dates of periodic calibrations (Holben et al. 1998). This interpolation technique is based on an assumption that F_0 (315 nm) constantly decreases with time. This assumption may not be perfectly valid (see Table 1 and Fig. 3). Never-

Fig. 3. $F_0(315 \text{ nm})$ values determined through Improved Langley (IL) technique described in Section 3.1 (original) and using Eq. (3) and data of different months. While using Eq. (3) and data of different months, a single value of OMI observation per day was assumed to be constant within 13:00–14:00 JST and estimated ozone concentrations from sky radiometer data within this period were used. The average values with standard deviations are shown in the figure. The given percentages in the figure are deviation percentages from the original value.

theless, it is the best option left at present to capture the temporal change of $F_0(315 \text{ nm})$.

5. Uncertainty analysis

In Eq. (2), the possible candidates to influence the estimated U can be considered to be τ_{aerosol} , F , F_0 , and K_{ozone} . We discuss how the uncertainty in each parameter is associated with the estimated U by assuming that the values of remaining parameters are unchanged.

For the uncertainties of $\Delta\tau_{\text{aerosol}}$ in τ_{aerosol} , ΔF in F , ΔF_0 in F_0 , and ΔK_{ozone} in K_{ozone} , the uncertainties in the estimated U denoted by ΔU_1 , ΔU_2 , ΔU_3 , and ΔU_4 , respectively, can be expressed as

$$\Delta U_1 = \frac{-1}{m_{\text{ozone}} K_{\text{ozone}}} \Delta \tau_{\text{aerosol}} m_{\text{air}}, \quad (4)$$

$$\Delta U_2 = \frac{-1}{m_{\text{ozone}} K_{\text{ozone}}} \ln\left(1 + \frac{\Delta F}{F}\right), \quad (5)$$

$$\Delta U_3 = \frac{1}{m_{\text{ozone}} K_{\text{ozone}}} \ln\left(1 + \frac{\Delta F_0}{F_0}\right), \quad (6)$$

and

$$\Delta U_4 = -\frac{\Delta K_{\text{ozone}}}{K_{\text{ozone}} + \Delta K_{\text{ozone}}} U. \quad (7)$$

The simplicity of Eqs. (4)–(7) allows one to easily evaluate the uncertainties in U due to the uncertainties in the input parameters. The values corresponding to different ranges of uncertainties in the input parameters are given in Fig. 4. Data corresponding to only positive values of the uncertainties in the input parameters are given in Fig. 4. Eq. (4) suggests that if τ_{aerosol} is overestimated, U is underestimated and vice versa. The uncertainties in U at different solar zenith angles (SZA) corresponding to different ranges of uncertainties in τ_{aerosol} are shown in Fig. 4a. As shown in Fig. 4a, one can roughly interpret ΔU_1 as $-\Delta\tau_{\text{aerosol}}/k_{\text{ozone}}$, if SZA is not very high. On the other hand, Eqs. (5) and (6) suggest that the uncertainty in U corresponding to the uncertainty in F or F_0 strongly depends on SZA. An overestimated F underestimates U , whereas an overestimated F_0 also overestimates U . Eqs. (5) and (6), as well as Figs. 4b and 4c suggest

Fig. 4. Uncertainties in columnar ozone concentration (U) as a function of solar zenith angle for different ranges of uncertainties in (a) aerosol optical thickness at 315 nm ($\Delta\tau_{\text{aerosol}}$), (b) direct intensity at 315 nm (ΔF), and (c) the calibration constant for direct intensity at 315 nm (ΔF_0), which are denoted by ΔU_1 , ΔU_2 and ΔU_3 respectively, and (d) the relative uncertainty in U corresponding to different ranges of relative uncertainties in ozone absorption coefficient at 315 nm (K_{ozone}). Those ranges of uncertainties are derived from Eqs. (4)–(7). Data for only positive values of uncertainties in the input parameters are given in the figure.

that the same uncertainty range in either F or F_0 can be less effective in the estimated U in the morning and afternoon rather than around noon. As further suggested by Figs. 4b and 4c, the uncertainty in F_0 (or F) should be less than 1 % to estimate U with less than 0.02 atm-cm uncertainty. Unlike other equations, Eq. (7) does not suggest the effect of SZA on the uncertainty in U due to the uncertainty in K_{ozone} . In Fig. 4d, we show the uncertainty in U due to the uncertainty in K_{ozone} in terms of relative values for the different ranges of uncertainties in K_{ozone} . For

example, if the value of K_{ozone} is larger than its true value by 5 %, it may underestimate the estimated U by roughly the same percentage.

6. Conclusions

A method to estimate U using the 315-nm-channel data of the sky radiometer is presented, and the preliminary results obtained from the proposed method are validated using the OMI observed data. As an alternative option of the traditional Langley method, a technique of determining F_0 (315 nm) using

in situ observation data is proposed. The relatively good agreement between the sky radiometer and OMI observed values for approximately two months and a gradual overestimation from the sky radiometer during the later observation period were noted. We speculate that such overestimated values from the sky radiometer are due to the temporal change of F_0 (315 nm), which was not possible to consider in the present study. This speculation is supported by the observation of the gradual decreases of F_0 values at the standard wavelengths used for studying the aerosol properties with the sky radiometer. The overall conclusion of this study is that the proposed method may be applicable to estimate U from the 315-nm-channel data of the sky radiometer, if F_0 (315 nm) is periodically determined.

This study recognizes the temporal change of F_0 (315 nm) as the major challenge for using long-term observation data of the sky radiometer to estimate U values. Future studies may find logical solutions to this problem. At present, the observations of reliable sky radiances at 315 nm are still a challenge due to associated technical problems. In the future, such data may be available and the determination of the temporal change of F_0 (315 nm) may be possible like other standard channels for aerosol observation. Until such progress is made, F_0 (315 nm) may need to be periodically determined using the procedure described in Section 3.1 to bring the data of the 315 nm channel in use to estimate U values.

Through this study, we come to the conclusion that the sky radiometer has a potential to observe columnar ozone, along with other atmospheric parameters, such as aerosols, precipitable water content, and clouds, which can be further improved in the future by collecting more data and conducting detailed investigations.

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