

Response to comments of referee 1

Authors would like to express sincere thanks to the referee 1 for valuable comments. We revised a manuscript carefully based on given comments. The comments of the referee 1 are in blue, our replies are in black, and changes made in the revised manuscript are in red. The English in this document has been checked by at least two professional editors, both native speakers of English. Our replies to the comments are as below.

General comments

The manuscript by Momoi et al. describes a novel method to self-calibrate the POM sun/sky radiometer for water vapour (WV) retrieval using diffuse sky radiance measurements and to estimate precipitable WV from direct irradiance at 940 nm (using a more "physical" approach than a non-linear empirical parametrisation of the Bouguer-Lambert-Beer law). The method is thoroughly and clearly explained, and the description is supported by sensitivity tests using radiative transfer models. The manuscript is skewed in favour of a theoretical/modelling perspective, with only two ending paragraphs focussed on experimental data, which is however justifiable owing to the main purpose of presenting a new method rather than studying the retrieved dataset. I recommend the publication of the manuscript after the authors have addressed some minor remarks, mainly aimed at improving readability by the unexperienced reader.

Specific comments

1. The new method seems to provide slightly worse results compared to AERONET, although POM and Cimel instruments are similar. It would be good if the authors could elaborate on this, thus enhancing a bit the experimental part of the paper. What is the most likely reason for this result? Is it due to the more physical (less empirical) approach employed in the study, with fewer empirical constraints? Have the authors explored the sensitivity of the retrievals to the accuracy of the instrumental characterisation (e.g., filter response function, field of view, etc.), to the used spectroscopic data (cross sections) or vertical profiles? If so, they could mention some of their results. More generally, on the basis of what criteria can the results of the WV retrieval be considered satisfactory? What are the maximum expected/permissible deviations, using such kind of instrument?

2. At least one plot of the time evolution of the retrieved w should be presented, also in order to understand when the maximum deviations from reference instruments occur;

The DSRAD algorithm retrieved the precipitable water vapor by a physical approach although the previous study of the sky-radiometer used an empirical approach (Uchiyama *et al.*, 2014; Campanelli *et al.*, 2014, 2018). The AERONET also uses the empirical method (Holben *et al.*, 1998). Our retrieved value was slightly worse compared with AERONET retrievals and others. The underestimation of
40 $PWV_{DSRAD+SKYMAP}$ was due to two factors. The first is the retrieval of PWV by the annual mean calibration constant for the water vapor channel. The calibration constant not only is subject to aging but also undergoes seasonal variation due to temperature dependency (Uchiyama *et al.*, 2018a). Thus, it is possible to underestimate the calibration constant in the wet season. Second, uncertainty regarding the
45 aerosol optical thickness affected PWV retrieval. Figure 18 depicts the differences in PWV and aerosol optical thicknesses at 675, 870, and 1020 nm between the DSRAD algorithm and the AERONET retrieval. In the periods from January to May and from October to November, the differences in PWV and aerosol optical thicknesses were less than 0.1 cm and 0.015, respectively. However, the difference in PWV was greater than 0.1 cm from July to September. This corresponds to the difference in aerosol optical thicknesses at 675, 870, and 1020 nm from July to September, which indicates that the
50 transmittance of water vapor was overestimated by the overestimation of aerosol optical thickness. This led to the underestimation of $PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant when PWV was > 3 cm. The above description is added in the revised manuscript (L612-625) and we also added the time series of PWV in the revised manuscript (Figure 18).

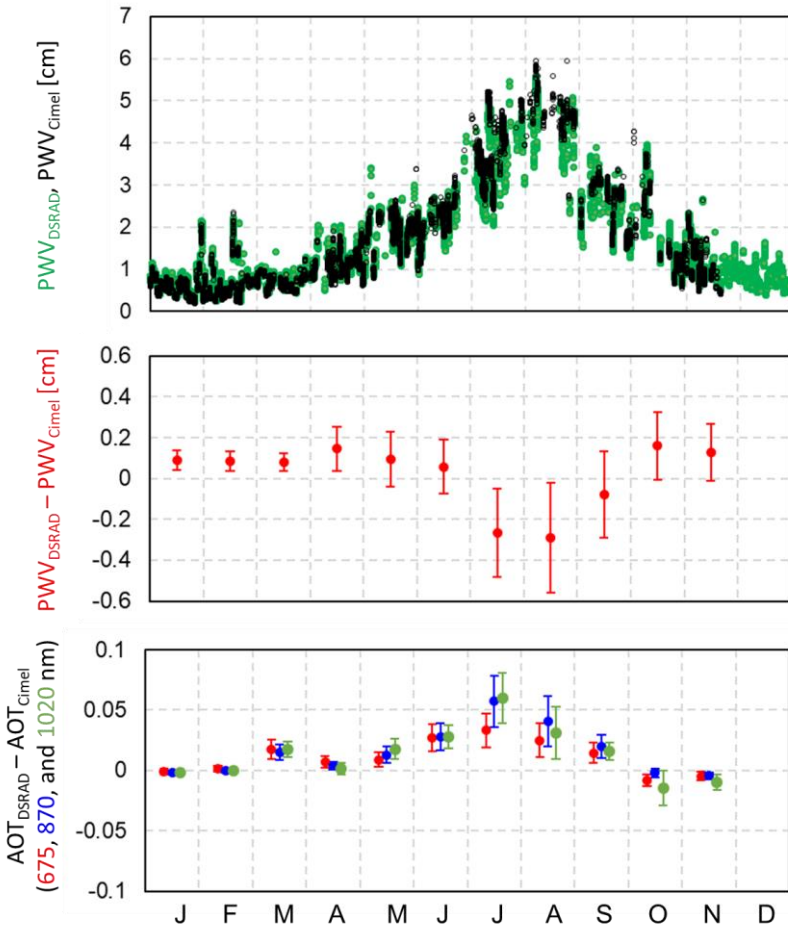


Figure 18: The top row shows the time series of PWV in 2017 at Chiba (green and black circles are $PWV_{DSRAD+SKYMAP}$ and PWV_{CIMEL} , respectively). The middle row is the difference between $PWV_{DSRAD+SKYMAP}$ and PWV_{CIMEL} . The bottom row is the difference in aerosol optical thicknesses at 675 nm (red), 870 nm (blue), and 1020 nm (green) between the DSRAD algorithm and the AERONET retrieval results. Circles and error bars in the middle and bottom rows are means and standard deviations, respectively.

3. It should be stressed that the sensitivity tests using synthetic data do not include measurement noise. If the authors also made some tests with noise, it would be interesting to present those results in the paper;

We conducted the sensitivity tests using the simulated data with the bias errors in the diffuse radiances. It is added in the revised manuscript (L510-521) as below:

We also conducted sensitivity tests using the simulated data with bias errors to investigate uncertainty in the SKYMAP-derived PWV. The bias errors were $\pm 5\%$ and $\pm 10\%$ for R . The value of 5% was given by following reasons. The SVA bias errors of the diffuse radiances for the sky-radiometer observations were estimated to be less than 5% (Uchiyama *et al.*, 2018b). According to Dubovik *et al.* (2000), the uncertainty of the diffuse radiances for the AERONET measurements is $\pm 5\%$. Figures 13 and 14 show

the results from the simulated data for the continental average and transported dust aerosols with aerosol optical thicknesses of 0.02, 0.06 and 0.20 at 940 nm. PWV was overestimated when -5% bias was applied to R . This corresponds to the relationship between R and PWV, where R decreases with increasing PWV (Section 2.1.2). The bias errors strongly affected the retrieval of PWV at high PWV (> 2 cm), because the sensitivity of high PWV is lower than that of low PWV. The retrieval error of PWV increased with increasing bias errors. The retrieval error of PWV due to $\pm 5\%$ and $\pm 10\%$ errors for R was within 10% for $\text{PWV} < 2$ cm and up to 200% for $\text{PWV} > 2$ cm.

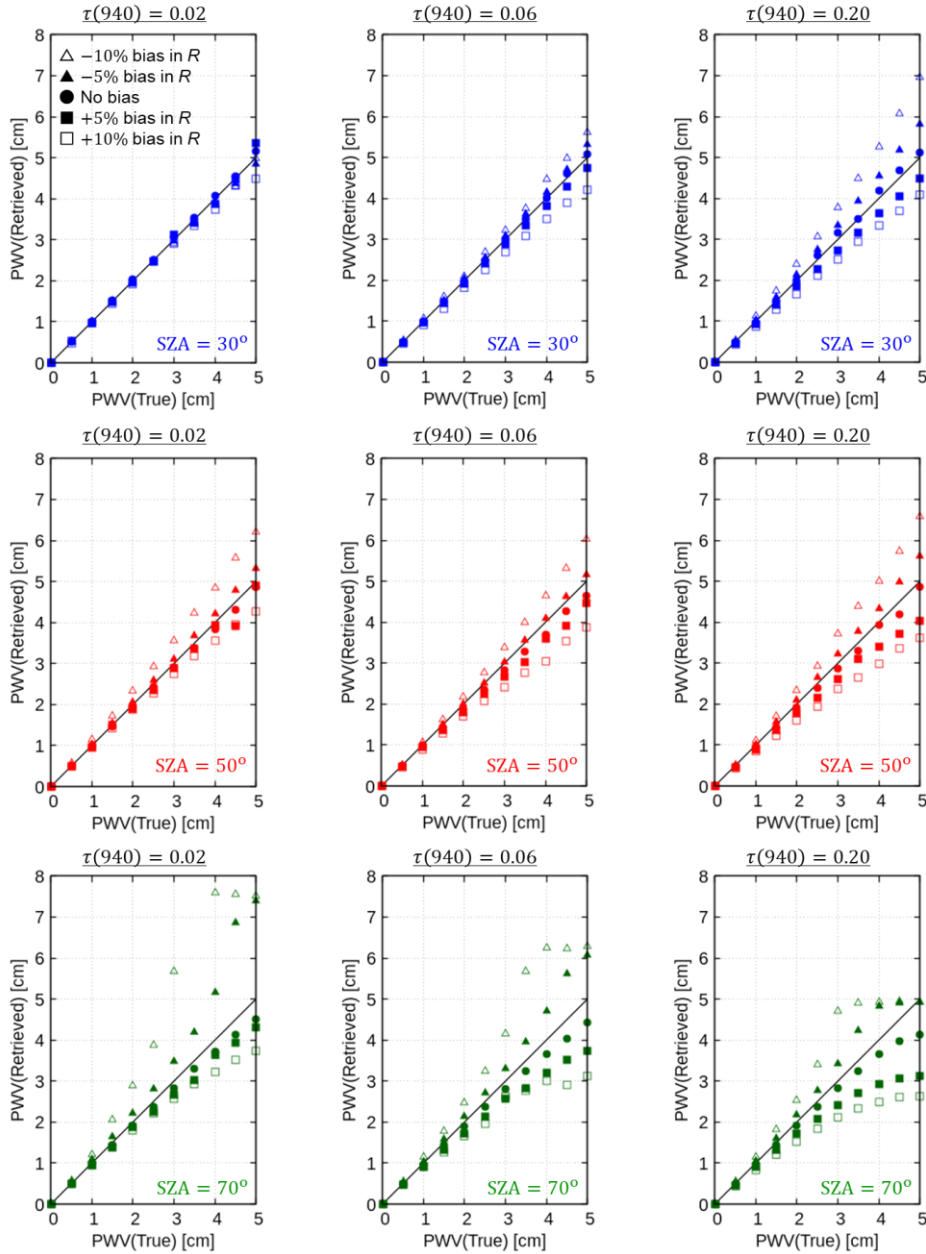


Figure 13: Comparison of the “true” and retrieval values of PWV from simulated data for continental average aerosol with bias errors. The top, middle, and bottom rows are the retrieval results at $\text{SZA} = 30^\circ$, 50° , and 70° , respectively. Closed circles are the results with no bias errors. Closed squares and closed triangles are the results with bias errors of plus and minus 5% in R ,

respectively. Open squares and open triangles are the results with bias errors of plus and minus 10% in R , respectively.

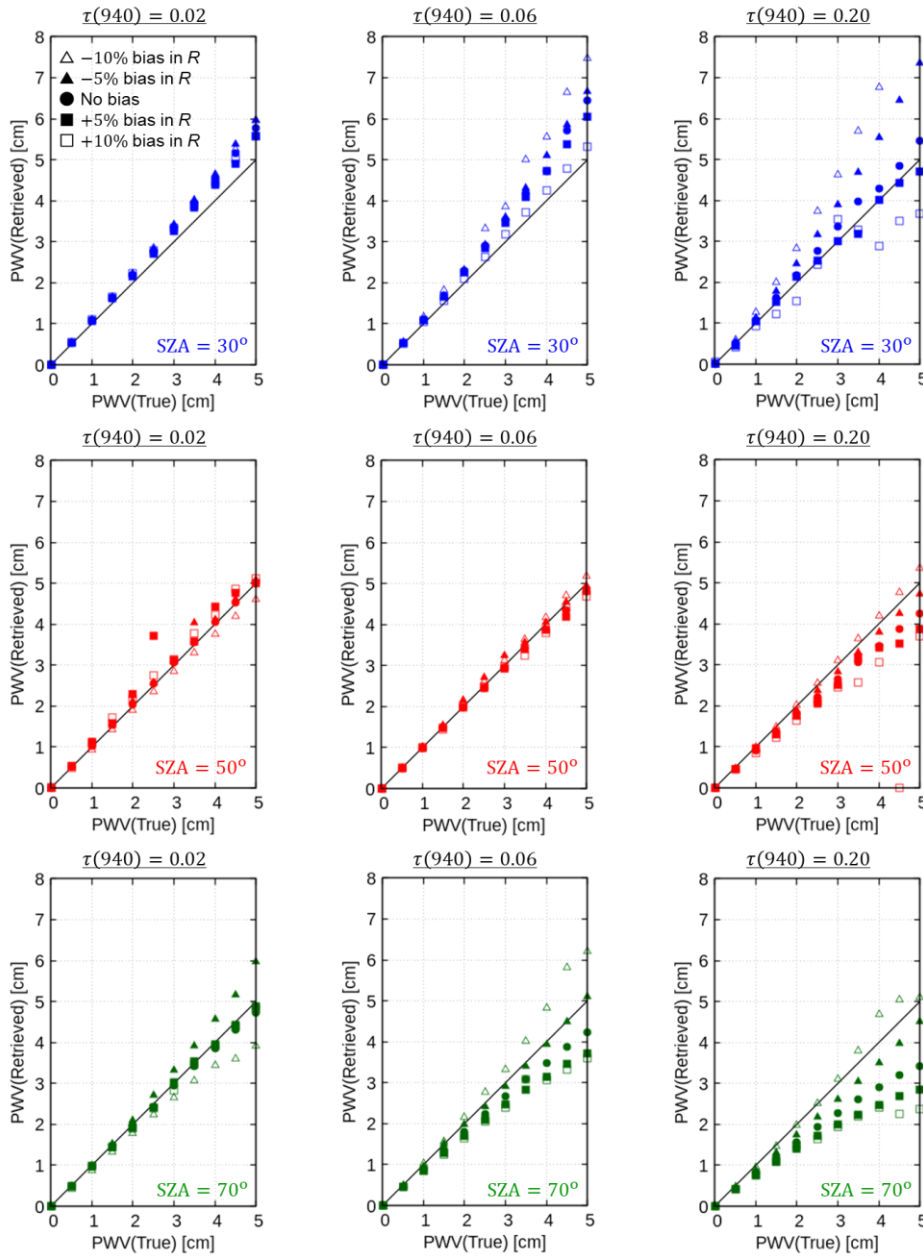


Figure 14: Similar to Fig. 13 but for transported dust aerosol.

We also compared our retrievals with more establish methods, such as GNSS/GPS-derived PWV, and AERONET-derived PWV as below:

95 (Section 4.1: L582-587)

Table 5 summarizes the results of comparisons of DSRAD-derived PWV and GNSS/GPS-derived PWV. The magnitude of the bias error and root mean square error were small, less than 0.11 cm and less than 0.226 cm, during 2013 to 2014. Table 6 shows the errors of the retrieved PWV with the annual mean calibration constants for the rank of PWV. The bias error was larger for high PWV than it was for low

100 PWV. The magnitude of the bias errors of PWV was less than 0.163 cm for $PWV < 3$ cm and less than 0.339 cm for $PWV > 3$ cm.

(Section 4.2: L602-610)

105 $PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant agreed with PWV_{MWR} (Fig. 17c). The error of $PWV_{DSRAD+SKYMAP}$ was $-0.041 < \text{bias} < 0.024$ cm and $RMSE < 0.212$ cm for low PWV (< 3 cm) and $\text{bias} < -0.356$ cm and $RMSE > 0.465$ cm for high PWV (Table 6). Figure 17d shows that $PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant also agreed with PWV_{Cimel} for low PWV (< 3 cm) but was smaller than PWV_{Cimel} for high PWV (> 3 cm). PWV_{MWR} was larger than PWV_{Cimel} (Fig. 17e). $PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant was 12% and 9.1% smaller than PWV_{MWR} and PWV_{Cimel} , respectively (Table 5).

Table 5: Comparison of PWV between DSRAD and other instruments.

		Slope C_1	Intercept C_2 [cm]	γ	Bias [cm]	RMSE [cm]
PS1202091 at Tsukuba, Japan						
Monthly mean F_0	vs GNSS/GPS receiver (2013)	0.956	0.079	0.938	-0.049	0.138
	vs GNSS/GPS receiver (2014)	0.937	0.161	0.970	-0.110	0.170
Annual mean F_0	vs GNSS/GPS receiver (2013)	0.919	0.173	0.987	-0.061	0.226
	vs GNSS/GPS receiver (2014)	0.934	0.178	0.987	-0.089	0.223
PS2501417 at Chiba, Japan						
Monthly mean F_0	vs MWR (2017)	0.964	0.053	0.961	-0.027	0.091
	vs AERONET (2017)	0.987	0.107	0.976	0.098	0.122
Annual mean F_0	vs MWR (2017)	0.880	0.132	0.985	0.042	0.231
	vs AERONET (2017)	0.909	0.184	0.991	0.055	0.186

C_1, C_2 : $PWV_{DSRAD} = C_1 \times PWV_{Other} + C_2$

Bias: $PWV_{DSRAD} - PWV_{Other}$

115 **Table 6: Difference in PWV between DSRAD with the annual mean calibration constants and other instruments.**

	PWV_{Other}				
	0 – 1 cm	1 – 2 cm	2 – 3 cm	3 – 4 cm	> 4 cm
	Bias [cm] (RMSE [cm])	Bias [cm] (RMSE [cm])	Bias [cm] (RMSE [cm])	Bias [cm] (RMSE [cm])	Bias [cm] (RMSE [cm])
PS1202091 at Tsukuba, Japan					
vs GNSS/GPS receiver (2013)	0.083 (0.124)	0.160 (0.211)	0.084 (0.236)	-0.098 (0.326)	-0.339 (0.537)
vs GNSS/GPS receiver (2014)	0.110 (0.142)	0.163 (0.221)	0.107 (0.251)	-0.055 (0.353)	-0.239 (0.492)
PS2501417 at Chiba, Japan					
vs MWR (2017)	0.017 (0.066)	0.024 (0.153)	-0.041 (0.212)	-0.356 (0.465)	-0.594 (0.722)
vs AERONET (2017)	0.088 (0.105)	0.118 (0.192)	0.017 (0.223)	-0.214 (0.386)	-0.264 (0.306)

Bias: $PWV_{DSRAD} - PWV_{Other}$

4. The present algorithm splits the instrumental characterisation (F0, bandpass, FOV) and the atmospheric parameters (WV profiles), while previous approaches use mixed empirical coefficients (a and b) dependent on both the spectral bandpass and the vertical WV profile. If my understanding is correct, this would permit to use the algorithm in different conditions (place/time) compared to the ones when the instrument was calibrated. Could this be an advantage to be underlined in the conclusions?

Yes, exactly. We add the sentence in the conclusion (L631-636) as below:

Our DSRAD algorithm retrieves PWV from the direct solar irradiance. This method does not require adjustment parameters used in the empirical methods of previous studies (e.g., Holben *et al.*, 1998; Uchiyama *et al.*, 2014; Campanelli *et al.*, 2014, 2018). Instead, the filter response function and the vertical profiles of water vapor, temperature, and pressure are required as input parameters. Thus, our physics-based algorithm has the potential to be applied to sky-radiometers all over the world. This is the greatest advantage of the present study.

Technical corrections

- l. 19, "whose aerosol channels": too abrupt beginning, especially for the readers not experienced in measurements with POM photometers. I would argue that aerosols are seldom the only influencing factor at a specific wavelength, therefore "aerosol channel" sounds more like a colloquial shortcut than a technical term;

- l. 28, "aerosol channels": specify the wavelengths;

140

We agree with the reviewer. We revised it (L19-26) as below:

The Prede sky-radiometer measures direct solar irradiance and the angular distribution of diffuse radiances at the ultraviolet, visible, and near-infrared wavelengths. These data are utilized for remote sensing of aerosols, water vapor, ozone, and clouds, but the calibration constant which is the sensor output current of the extraterrestrial solar irradiance at the mean distance between the Earth and the sun, is needed. The aerosol channels, which are the weak gas absorption wavelengths of 340, 380, 400, 500, 675, 870, and 1020 nm, can be calibrated by an on-site self-calibration method, the Improved Langley method.

- l. 21-22, "by sky-radiometer remain challenge": some articles (a/the) missing;

We agree with the reviewer. We revised it (L28) as below:

by the sky-radiometer remains challenging

- 155 - l. 22-23, "generally calibrated by the standard Langley method": only for reference instruments, then calibration is transferred to network instruments (as explained later in the text);
- l. 24, "water vapor channel": please mention the channel wavelength here;

We agree with the reviewer. We revised it (L27-31) as below:

- 160 However, the continuous long-term observation of precipitable water vapor (PWV) by the sky-radiometer remains challenging, because calibrating the water vapor absorption channel of 940 nm generally relies on the standard Langley method (SL) at limited observation sites (*e.g.*, the Mauna Loa Observatory) and the transfer of the calibration constant by side-by-side comparison with the reference sky-radiometer calibrated by the SL method.

165

- l. 53, "columns of water vapor content": do you mean column concentrations? Unclear, since radiosonde also measure vertical profiles;

It means "precipitable water vapor". We revised it (L76).

170

- l. 60, "these previous studies": even the studies by Fowle («1992)?

We intended "these previous studies" points "SKYNET sky-radiometer (Campanelli *et al.*, 2014, 2018; Uchiyama *et al.*, 2014, 2018a), and AERONET sun-sky photometer (Holben *et al.*, 1998)". Therefore,
175 we revised it (L84-87) as below:

Previous studies of SKYNET and AERONET derived PWV from the observed transmittance of water vapor ($\bar{T}_{\text{H}_2\text{O}}$), assuming $\bar{T}_{\text{H}_2\text{O}} = e^{-a(m \cdot w)^b}$ (Bruegge *et al.*, 1992), where a and b are adjustment parameters, m is the optical air mass, and w is PWV.

- 180 - l. 64: do you really mean "radiometric calibration" (as in F0) or, *e.g.*, "spectral sensitivity"?

We intend "spectral sensitivity of spectroradiometer". We revised it (L87-89) as below:

- However, there is a known noticeable uncertainty in the estimate of PWV because the adjustment parameters depend on the spectral sensitivity of the spectroradiometer as well as the vertical profiles of
185 water vapor and temperature.

- l. 75, "Sky-radiometer": article (the) missing?

We revised it (L102) as below:

- 190 The sky-radiometer models POM-01 and POM-02 (Prede, Tokyo, Japan), which are ...

- l. 77, "11 wavelengths": maybe a table of the channel wavelengths, together with the main extinction factors, could be useful. See also my first technical comment about the expressions "aerosol channels ... ozone channels" (e.g., even the "ozone channel" is affected by aerosol)

195 - l. 107-108, "aerosol ... cloud ... water vapor ... ozone channels": cf. previous comments. These approximate expressions could be used only after a short explanation;

We agree with reviewer. We revised it (L102-109) and added the table of sky-radiometer specifications as below:

200 The sky-radiometer models POM-01 and POM-02 (Prede, Tokyo, Japan), which are deployed in the international radiation observation network SKYNET, measure solar direct irradiances and diffuse irradiances at the ultraviolet, visible, and near-infrared wavelengths. These measurements are used for the remote sensing of aerosol, cloud, water vapor, and ozone (Table 1; Takamura and Nakajima, 2004; Nakajima *et al.*, 2007). Table 1 shows the relationship between the wavelengths and the main target of
205 the remote sensing. The aerosol channels are 340, 380, 400, 500, 675, 870, and 1020 nm; the water vapor channel is 940 nm; the ozone channel is 315 nm; and the cloud channels are 1225, 1627, and 2200 nm.

210 **Table 1: Sky-radiometer specifications. Each sky-radiometer is equipped with a filter indicated by a circle. "Standard" is the standard specification of sky-radiometer models POM-01 and POM-02.**

Wavelength [nm]	Strong gas absorption	Main target substance	POM-01 Standard	POM-02 Standard	POM-02 PS1202091	POM-02 PS2501417
315	O ₃	Ozone	○	○	—	○
340	—	Aerosol	—	○	○	○
380	—	Aerosol	—	○	○	○
400	—	Aerosol	○	○	○	○
500	—	Aerosol	○	○	○	○
675	—	Aerosol	○	○	○	○
870	—	Aerosol	○	○	○	○
940	H ₂ O	Water vapor	○	○	○	○
1020	—	Aerosol	○	○	○	○
1225	O ₂ , CO ₂ , H ₂ O	Cloud	—	—	○	—
1627	CH ₄ , CO ₂	Cloud	—	○	○	○
2200	CH ₄ , H ₂ O	Cloud	—	○	○	○

- l. 79, "observation ... self-calibration": a bit confusing, please reformulate to avoid mixing of observation and calibration procedures;

215 - l. 80, "works in turbid atmospheric conditions": "only" in turbid conditions or "also" in turbid conditions?

We agree with the reviewer. We revised it (L109-114) as below:

220 Through on-site self-calibration of the aerosol channels by the Improved Langley (IL) method (Tanaka *et al.*, 1986; Nakajima *et al.*, 1996; Campanelli *et al.*, 2004, 2007), the SKYNET system is capable of long-term and continuous aerosol observation. The IL method works not only in clean atmospheric conditions, but also in turbid atmospheric conditions.

- l. 81-82, "standard ... modified": please, define what a "standard" method and a "modified" one are;

225

The standard and modified Langley methods are developed by Uchiyama *et al.*, (2014) and Campanelli *et al.* (2014), respectively. We revised it (L115-116) as below:

However, no improved calibration method has replaced the standard (Uchiyama *et al.*, 2014) or modified (Campanelli *et al.*, 2014, 2018) Langley methods for the water vapor channel.

230

- l. 93, "two SKYNET sites": explain why these two sites were selected. Do they have any particular characteristics, or was this choice oriented by the co-located instrumentation?

We chose sites which had been installed not only the sky-radiometer, but also AERONET sun-sky
235 radiometer, GPS/GNSS receiver, and/or MWR. We added the sentence at the end of Section 1 (L129-130) as below:

At these two sites, PWV is observed by the GNSS/GPS receiver, MWR, or AERONET sun-sky radiometer other than the sky-radiometer.

240 - l. 103, "We explain normalized radiance": article missing?

We agree with the reviewer. We revised it (L142) as below:

We explain the normalized radiance...

245 - Eq. (1): specify earlier in the text that this holds only for a plane-parallel nonrefractive atmosphere (l. 122, now);

We agree with the reviewer. We revised it (L154-162) as below:

250 In the plane-parallel non-refractive atmosphere, F at the bottom of the atmosphere (BOA) at the solar zenith angle (SZA) θ_0 and the solar azimuth angle ϕ_0 is derived from

$$F(\lambda) = \frac{F_0}{d^2} \exp(-m_0 \tau(\lambda)), \quad (1)$$

where F_0 is the calibration constant; d is the distance between Earth and the sun (AU); λ is the wavelength; τ is the total optical thickness; and m_0 is optical air mass, represented as $m_0 = 1/\cos\theta_0$.

- Eq. (2): if L is defined as sky radiance (l. 106), then it should be already divided by the solid view angle (ω);

We used the sky irradiance instead of the sky radiance in the revised manuscript.

- l. 150: the sentence is missing its subject. It is also unclear if this limitation (the real atmosphere not being a single layer) will be addressed in the following text;

We agree with the reviewer. It is deleted and mentioned in Sect. 2.1.2 in the revised manuscript.

- l. 152, "sensitivity of R ": ... at a wavelength of 940 nm;

Yes, it is at a wavelength of 940 nm. We revised it (L193) as below:

We examined the sensitivity of R at 940 nm ...

- l. 154: I guess that these AOD values refer to 940 nm, too?

Yes exactly. We revised it (L199-201) as below:

Figure 3 shows the dependencies of R in the almucantar plane on PWV for continental average aerosol with aerosol optical thicknesses of 0.02 and 0.20 at 940 nm.

- l. 155-156, "the aerosol optical thickness does not affect this relationship": unclear, since R decreases, but the values do depend on AOD;

We agree with the reviewer. The relationship that R decreases with an increase of PWV was seen in both low and high AOT cases. We revised it (L201-203) as below:

R decreases with increasing PWV regardless of the aerosol optical thickness.

- l. 164: I would not define such a change as "drastic". The variation is only visible in the lower subfigures with a linear y-axis (please, put some letters next to the subplots), and mainly for $PWV \leq 2$ cm (as explained later in the text);

It is deleted the word “drastic” in the revised manuscript (L212).

290

- l. 188: is "SKYMAP" an acronym?

No, it isn't.

295 - l. 197, "transmittance of the total extinction": isn't just "transmittance" enough?

We agree with the reviewer. We revised it (L248).

- Eq. (8): please, explain where the 1.65 factor comes from;

300

This factor, which is defined as η in the revised manuscript, was determined in consideration of the smoothness of the VSD. It is written in the Appendix A in the revised manuscript.

Appendix A: Width of the volume size distribution

Because $\frac{dV(r)}{d\ln r}$ is expressed by the superposition of 20-modal lognormal size distributions (Eq. [6]), the
305 width of $\frac{dV(r)}{d\ln r}$ is larger than that of each lognormal size distribution. The width of the lognormal size
distribution should be small to deal with the complicated and step variations in $\frac{dV(r)}{d\ln r}$. However, $\frac{dV(r)}{d\ln r}$
cannot represent a natural curve if η is large and s is small (Fig. A1). Hence, we have to find the
maximum value of η for making $\frac{dV(r)}{d\ln r}$ a natural curve. When C_i is constant, such value of η minimizes
the roughness of $\frac{dV(r)}{d\ln r}$, and $\frac{dV(r)}{d\ln r}$ approaches to a flat shape. For a simple formulation, we consider the
310 function $A(x)$ which consists of the multimodal normal distribution function B_i with a constant height.
 $A(x)$ and B_i are expressed as

$$A(x) = \sum_{i=-\infty}^{\infty} B_i(x) = \sum_{i=-\infty}^{\infty} \exp \left[-\frac{\eta^2}{2} \left(\frac{x-i\xi}{\xi} \right)^2 \right], \quad (\text{A1})$$

315 where $i\xi$ and $\frac{\xi}{\eta}$ are the mean and standard deviation, respectively. Its differential is written as

$$\frac{dA}{dx} = \sum_{i=-\infty}^{\infty} \frac{dB_i}{dx} = \sum_{i=-\infty}^{\infty} -\eta^2 \left(\frac{x-i\xi}{\xi} \right) \exp \left[-\frac{\eta^2}{2} \left(\frac{x-i\xi}{\xi} \right)^2 \right]. \quad (\text{A2})$$

When the shape of $A(x)$ approaches to be flat, the difference between local maximum and minimum values of $A(x)$ is approximately 0. Because $\frac{dB_i}{dx}$ equals 0 at $x = j\xi$ ($j \in \mathbb{Z}$), $A(x)$ has the local maximum and minimum at $x = j\xi$ and $(j + \frac{1}{2})\xi$ in $j \leq \frac{x}{\xi} < j + 1$. The difference Δ between the local maximum and minimum values is obtained as

$$\Delta = 1 - \frac{A(\frac{2j+1}{2}\xi)}{A(j\xi)}. \quad (\text{A3})$$

325

Figure A2 shows the relation between η and Δ . The value of Δ increases drastically at around $\eta = 1.5$. in addition, the shape of $\frac{dV(r)}{d\ln r}$ is unnatural when $\eta = 2.0$ (Fig. A1). Therefore, the value of η should be selected from the values around $\eta = 1.5$. In this study, we fixed η at 1.65. This value represents the natural curve of $\frac{dV(r)}{d\ln r}$ and satisfies that the value of Δ is small enough, $\Delta = 3.0 \times 10^{-3}$.

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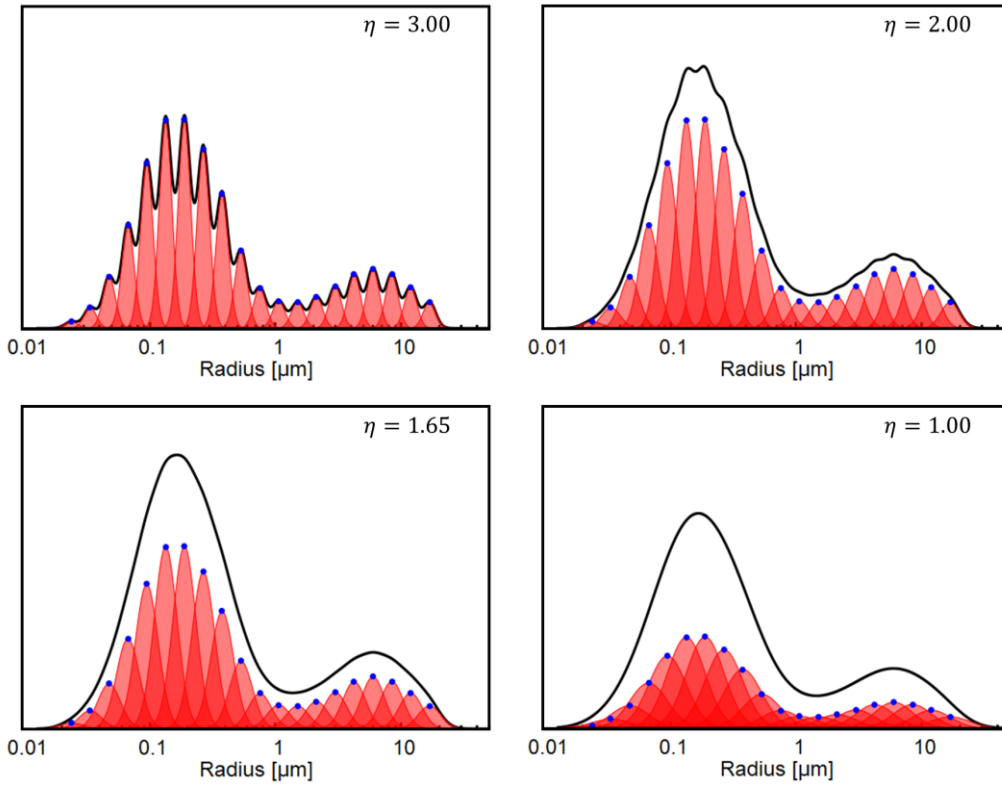


Figure A1: Relationship between the volume size distribution and η . The black line is the volume size distribution, which is computed by the integration of 20-modal lognormal distribution functions (red lines). Blue circles are the peak volume of lognormal size distribution.

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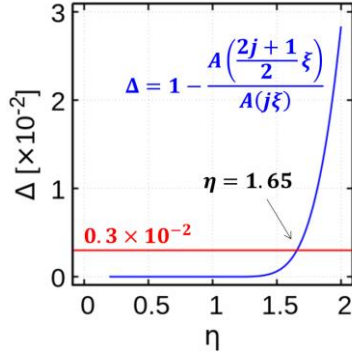


Figure A2: Relationship between the parameter η and the difference Δ .

340 - 1. 218, "local minimum": does this local minimum change at every retrieval, then?

Yes, exactly. The local minimum of the VSD is calculated at every retrieval.

345 - 1. 338: specify that the integral of the filter response function was normalised to 1 (not its maximum);

The response function is not normalized to 1. We revised it (L386) as below:

$$\bar{T}_{\text{H}_2\text{O}} = \frac{\int_{\Delta\lambda} \Phi(\lambda) T_{\text{H}_2\text{O}}(\lambda) d\lambda}{\int_{\Delta\lambda} \Phi(\lambda) d\lambda} = \frac{\int_{\Delta\lambda} \Phi(\lambda) \exp(-m_{\text{H}_2\text{O}}(\theta) \int_0^z \alpha_{\text{H}_2\text{O}}(g_w(z), K(z), \lambda) dz) d\lambda}{\int_{\Delta\lambda} \Phi(\lambda) d\lambda} \quad (22),$$

350 - Eq. (25)-(26): it could be useful to use a subscript (j?) for the single F_0 's. Also, please use another letter instead of w (in w_H);

We agree with the reviewer. We changed it (L408-417) as below:

The mean value of the calibration constant at the water vapor channel is determined by the robust statistical and iterative method with Huber's M-estimation:

355

$$\ln \bar{F}_0 = \sum_i \beta_H(t_i) \cdot \ln F_0(t_i), \quad (25)$$

$$\beta_H(t_i) = \begin{cases} 1 & (|\ln \bar{F}_0 - \ln F_0(t_i)| \leq 0.03) \\ \frac{0.03}{|\ln \bar{F}_0 - \ln F_0(t_i)|} & (|\ln \bar{F}_0 - \ln F_0(t_i)| > 0.03) \end{cases}, \quad (26)$$

360

where \bar{F}_0 is the mean calibration constant and is calculated at each iterative step, $F_0(t_i)$ is the calibration constant at a specific time t , and β_H is Huber's weight function.

- Eq. (28)-(29): perhaps it would be better to identify the indices with other letters than R (already used for radiances). Also, it should be mentioned that the range of scattering angles for the calculation of index 2 changes during the day;

We agree with the reviewer. We revised it (L430-442) as below:

Next, the running mean of the time series of $\bar{R}_{\text{near}}(t)$ with a window of three consecutive data points is calculated as $\langle \bar{R}_{\text{near}}(t) \rangle$. Index 1 is defined as the deviation $\tilde{R}_{\text{near}}(t)$ of $\bar{R}_{\text{near}}(t)$ from $\langle \bar{R}_{\text{near}}(t) \rangle$,

$$\tilde{R}_{\text{near}}(t) = |\bar{R}_{\text{near}}(t) - \langle \bar{R}_{\text{near}}(t) \rangle| / \langle \bar{R}_{\text{near}}(t) \rangle. \quad (28)$$

Index 2 is the deviation \tilde{R}_{far} of normalized angular distributions far from the sun and is defined as

$$\tilde{R}_{\text{far}}(t) = \sigma \left(\frac{R(\Theta, t) - \langle R_{\text{far}}(\Theta, t) \rangle}{\langle R_{\text{far}}(\Theta, t) \rangle} \right), \Theta > 10^\circ, \quad (29)$$

where $\langle R_{\text{far}}(\Theta, t) \rangle$ is the running mean of $R(\Theta_i, t)$ with a window of three consecutive data points, and $\sigma(\mathbf{X})$ is the standard deviation of data set \mathbf{X} . Note that the data for calculating \tilde{R}_{far} varies depending on SZA, which limits available scattering angles.

- l. 390, l. 397 and Table 2, "misjudged"/"incorrect": it is unclear whether the cloud screening criterium correctly works for the portion of the sky seen by the photometer or not. In the first case, the algorithm does its job, and I think that "misjudged"/"incorrect" are misleading terms, since the conditions of whole sky should not be considered as reference;

"misjudged" and "incorrect" were revised. And "clear-sky" and "cloud affected" were changed to "best condition" and "poor condition" in the revised manuscript (L461-463).

- l. 396-397: are <1 and >2 oktas?

It means cloud cover the range from 0 (no cloud) to 10 (cloud). We used the percentage of cloud cover for whole sky instead of previous "cloud cover the range from 0 (no cloud) to 10 (cloud)". We revised it.

- l. 407, "line regression": do you mean linear regression using AOD and wavelength (not logs)?

"line regression" means "linear interpolation in the log-log plane". We revised it (L473).

- Sect. 3: were only "aerosol channels" (l. 107) + "water vapor channel" used in the synthetic retrieval?

400

Yes, Sect. 3 conducted intensive sensitivity tests using aerosol channels and the water vapor channel.

- l. 464-465: are the -10% and -3% deviations within the expected uncertainty? If not, can you explain these results? Please, use a proper number of significant digits;

405

We consider that the difference in the value of the calibration constant between the SKYMAP algorithm and the side-by-side comparison with the reference sky-radiometer was attributable mainly to the calibration period. The calibration constant of the sky-radiometer has seasonal variation due to the temperature dependency of the sensor output (Uchiyama *et al.*, 2018a). Calibration by side-by-side
410 comparison with the reference sky-radiometer was performed only in the winter. However, the calibration constant of the SKYMAP algorithm was the annual mean. The above description is added in the revised manuscript (L551-556).

- l. 472: maybe it would be more scientifically correct to plot these values anyway (with another
415 colour/marker) even though they will not be considered;

We calculated the monthly mean calibration constant in all of observation periods (Fig. 15a, 16a, 17a). However, we did not retrieve PWV using the monthly mean calibration constants for June and July 2014 because their values were obviously small, and because little data were successfully retrieved due
420 to the wet and cloudy conditions in the summer. In addition, it is possible that the measurements were contaminated by clouds. The above description is added in the revised manuscript (L565-568).

- l. 478: a more natural choice would be to linearly interpolate the monthly calibrations. The authors certainly have good reasons for considering the annual mean value, can they elaborate on this?

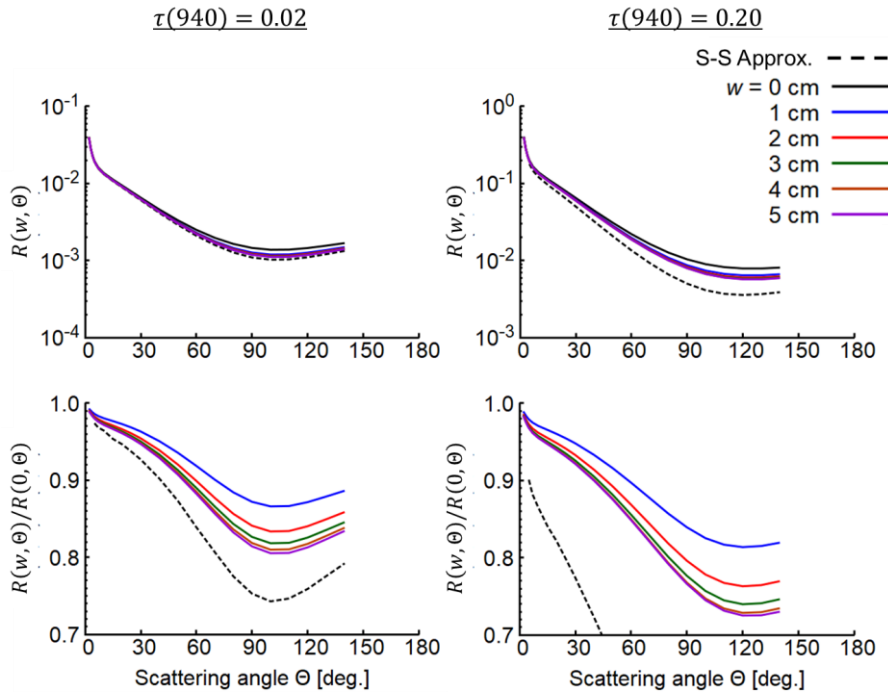
425

In this study, we used the annual mean calibration constant due to two factors. First, the monthly mean calibration constants were not significant changed in the dry seasons. Second, the accuracy of the calibration constant decreases at the transition between the wet season and the dry season because little data were successfully retrieved due to the wet and cloudy conditions.

430

- l. 518, "much more": "larger"?

Yes. We revised it (L639-640) as below:



455 **Figure 3: Normalized angular distributions simulated for continental average aerosol (Table 2) in the almucantar plane with aerosol optical thicknesses of 0.02 and 0.20 at 940 nm. Simulations were conducted for SZA = 70° and PWV (w) = 0, 1, 2, 3, 4, and 5 cm. The top row is the normalized radiance $R(w, \Theta)$, and the bottom row is the ratio of $R(w, \Theta)$ to $R(0, \Theta)$. S-S Approx. is single scattering approximation.**

460

- Fig. 10: colours in the second row are hardly distinguishable. Explain that they overlap (in the caption);

We agree with the reviewer. We added it in the caption of Figs. 11 and 12 in the revised manuscript as

465

below:

Note that the blue, red, green, and black lines in the middle row overlap.

- Table 3, "Retrieved the PWV": "PWV retrieval"?

470

We agree with the reviewer. We revised it as below:

Table 4: References and methodologies of the DSRAD algorithm.

DSRAD	
Solar coordinates	Nagasawa (1999)
Refraction correction	Nagasawa (1999)
Sun-Earth distance	Nagasawa (1999)
Optical mass	Gueymard (2001)
Rayleigh scattering	Fröhlich and Shaw (1980); Young(1981)
Ozone absorption	Sekiguchi and Nakajima (2008)
Water vapor absorption	Sekiguchi and Nakajima (2008)
Filter response function	Stepwise function
Retrieval of PWV	Newton-Raphson method

Response to comments of referee 2

Authors would like to express sincere thanks to the referee 2 for valuable comments. We revised a manuscript carefully based on given comments. The comments of the referee 2 are in blue, our replies are in black, and changes made in the revised manuscript are in red. The English in this document has been checked by at least two professional editors, both native speakers of English. Our replies to the comments are as below.

The study performed by Momoi et al., is a very important evolution in PWV retrieved from radiometers. It exploits findings of established methods and proposes an approach that can be used without prior calibration and could provide reliable PWV from sky-sun measurements. The methodology is described in detail and justified in an appropriate manner. Results are promising and the method could be adopted operationally from Skynet and other networks. I suggest accepting the manuscript for publishing in AMT, after some minor technical corrections and clarifications.

Since it is a novel method, it is crucial to add some preliminary discussion on the uncertainties of the method. It is important for scientists to have an estimation of the expected uncertainties, before applying the method. Is it more accurate than the well-established methods or the main advantage is the field calibration? Also, when transferring the calibration from the one method to the other, to retrieve PWV from the direct sun data, the error propagation is expected to be very high. I suggest discussing this issue in detail.

The main advantage of this manuscript is the field calibration of the water vapor channel (around 940 nm) of the sky-radiometer. In general, the transferring by side-by-side comparison increase the error, but our SKYMAP algorithm is self-calibrating and does not require transfer. The uncertainty of retrieved PWV is discussed by the simulated data with the bias errors in the diffused radiances in Section 3 (L510-521) in revised manuscript as below:

We also conducted sensitivity tests using the simulated data with bias errors to investigate uncertainty in the SKYMAP-derived PWV. The bias errors were $\pm 5\%$ and $\pm 10\%$ for R . The value of 5% was given by following reasons. The SVA bias errors of the diffuse radiances for the sky-radiometer observations were estimated to be less than 5% (Uchiyama *et al.*, 2018b). According to Dubovik *et al.* (2000), the uncertainty of the diffuse radiances for the AERONET measurements is $\pm 5\%$. Figures 13 and 14 show the results from the simulated data for the continental average and transported dust aerosols with aerosol optical thicknesses of 0.02, 0.06 and 0.20 at 940 nm. PWV was overestimated when -5% bias was applied to R . This corresponds to the relationship between R and PWV, where R decreases with

increasing PWV (Section 2.1.2). The bias errors strongly affected the retrieval of PWV at high PWV (> 2 cm), because the sensitivity of high PWV is lower than that of low PWV. The retrieval error of PWV increased with increasing bias errors. The retrieval error of PWV due to $\pm 5\%$ and $\pm 10\%$ errors for R was within 10% for $\text{PWV} < 2$ cm and up to 200% for $\text{PWV} > 2$ cm.

40

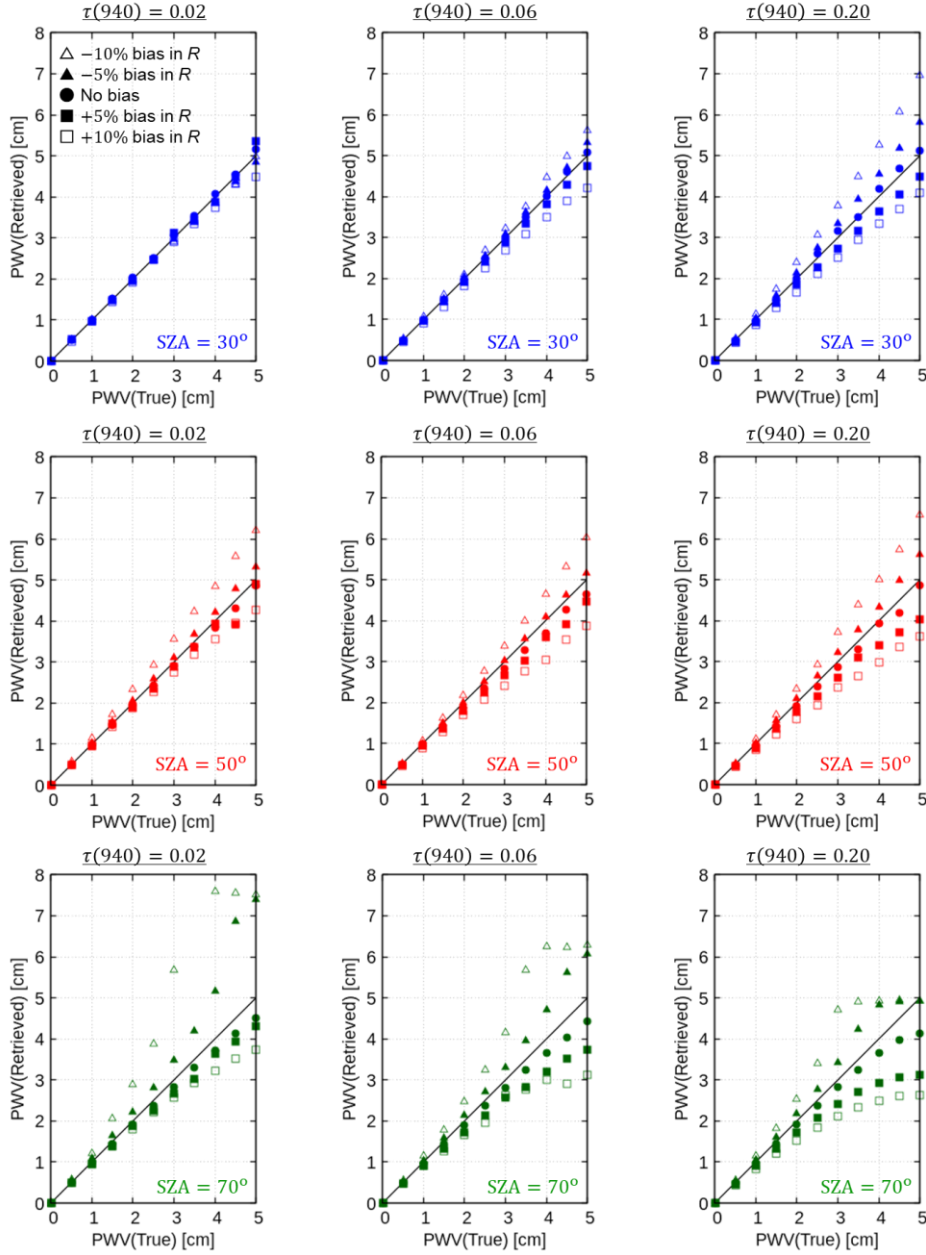
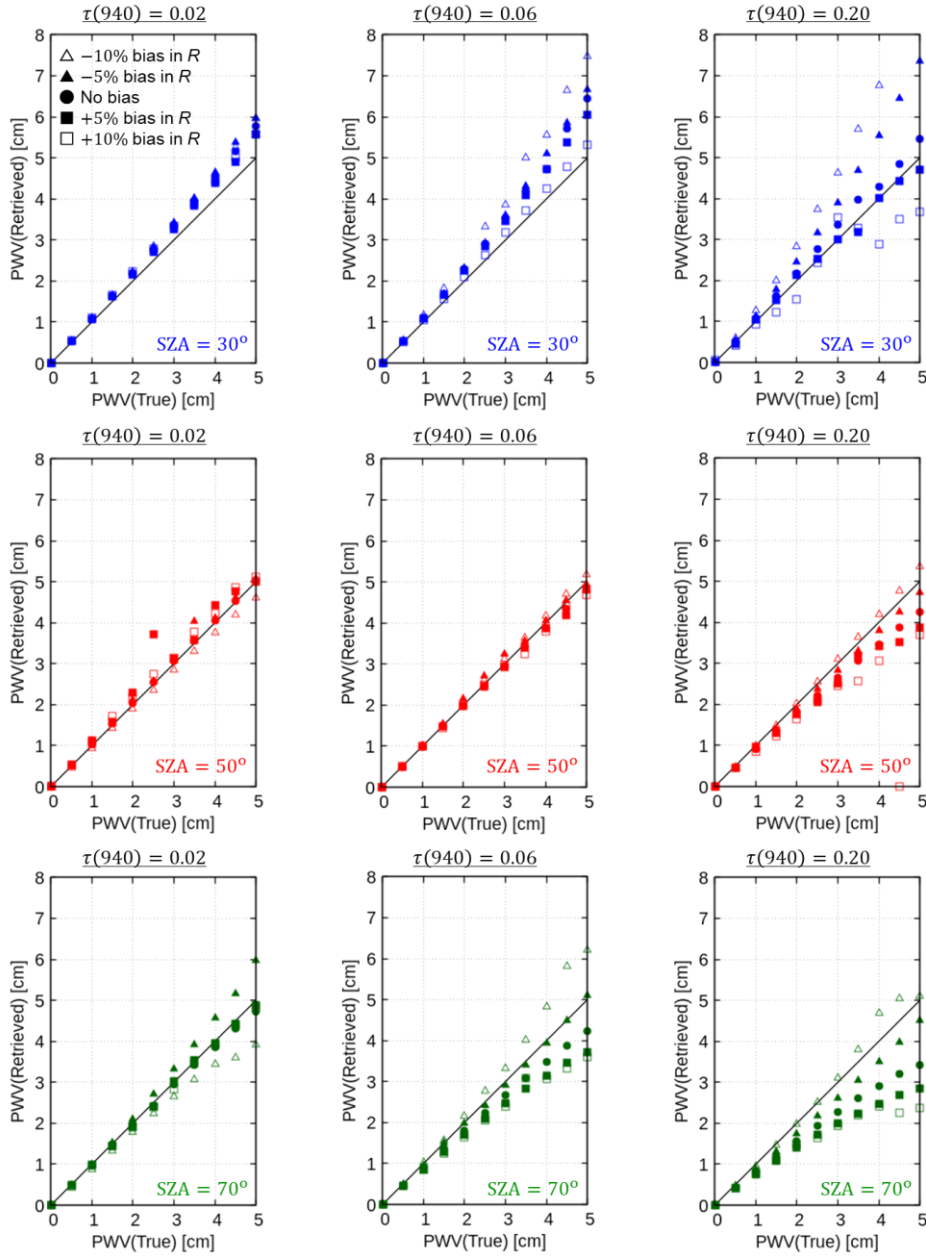


Figure 13: Comparison of the “true” and retrieval values of PWV from simulated data for continental average aerosol with bias errors. The top, middle, and bottom rows are the retrieval results at $\text{SZA} = 30^\circ$, 50° , and 70° , respectively. Closed circles are the results with no bias errors. Closed squares and closed triangles are the results with bias errors of plus and minus 5% in R , respectively. Open squares and open triangles are the results with bias errors of plus and minus 10% in R , respectively.

45



50 **Figure 14: Similar to Fig. 13 but for transported dust aerosol.**

It should be clarified in abstract and introduction section, that the calibration constant is referring to the extraterrestrial / Top of the atmosphere value of F . It might be reasonable for people into sunphotometry, but when referring to various equations and approaches, it should be crystal clear the referred quantity.

55

We agree with the reviewer. We added the explanation of the calibration constant in abstract and Section 1 in revised manuscript as below:

(Abstract: L20-23)

These data are utilized for remote sensing of aerosols, water vapor, ozone, and clouds, but the calibration constant which is the sensor output current of the extraterrestrial solar irradiance at the mean distance between the Earth and the sun, is needed.

60

To estimate PWV using a spectroradiometer, it is necessary to calibrate the water vapor channel. The calibration constant, which is the sensor output current of the extraterrestrial solar irradiance at the mean distance between the Earth and the sun, at the water vapor channel can be determined by the Langley method.

Method for cloud screening described in 2.2.4, is first time described? If it is not, or if it is based on an existing method, some references should be provided. If it totally new, it should be discussed more. Not just one day as an example for the validity. What are the improvements compared to other approaches? If 17% of cloud cases are contaminating the clear sky data, why don't change to different threshold values? 17% seem a big number which will end in high errors to the data set. At least for the validation of PWV method, stricter criterion is preferable, though it might result in smaller database, because the goal is to estimate the results of algorithm in clear sky conditions.

The SCAD method is newly developed in this study for the application to the observational site where only the sky-radiometer is installed. We add the further discussion about determining thresholds in the revised manuscript (L455-464) as below:

The results were validated using visual observation of the amount of clouds in the Aerological Observatory of the JMA. Figure 10a shows the histograms of index 1 for cases in which the sun was and was not covered by clouds. Index 1 had a low value when there were no clouds shading the sun but had a wide range of values when clouds were shading the sun. Fig. 10b shows the histograms of index 2 when cloud cover was and was not $< 20\%$. The peak shifted to the right when cloud cover was $\geq 20\%$, but the effect was not significant. Table 3 shows the validation results of this method. We defined "best condition" as cloud cover $< 20\%$ and "poor condition" as cloud cover $\geq 20\%$. In less than 17% of cases a "poor condition" was judged as a "best condition". The sky-radiometer observes only a part of the whole sky, but our algorithm showed good results.

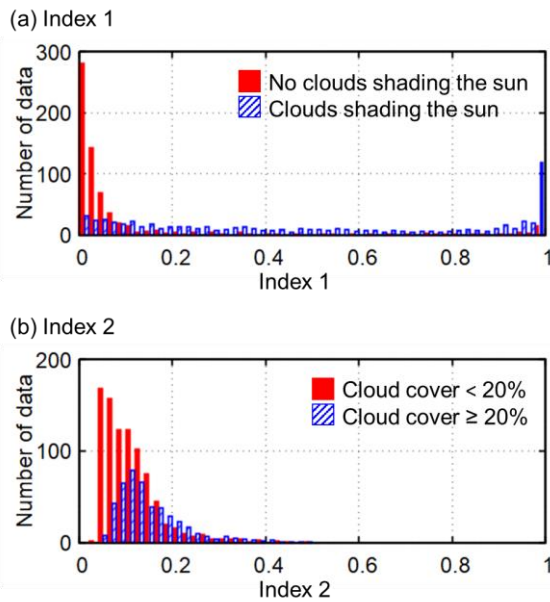


Figure 10: Histograms of indexes 1 and 2 of sky-radiometer observations at Tsukuba. (a) Index 1 when the sun is covered by clouds (blue boxes) and not covered by clouds (red boxes). (b) Index 2 when cloud cover is less than to 20% (red boxes) and greater than or equal to 20% (blue boxes).

Table 3: Validation of the SCAD method by visual observation from 2013 to 2014 in Tsukuba.

Visual observation Cloud cover	Sky-radiometer measuring plane	
	Available condition	Poor quality condition
Clear, less than 20%	463 (83.4%)*	68 (8.7%)
Cloud affected, more than 20%	92 (16.6%)	714(91.3%)*

*Obviously correct determination.

In step 2 of SKYMAP. Does it retrieves PWV or just the corresponding transmittance (as stated in the abstract?) If it is just the transmittance is should be clarified in the description and change the title of 2.2.2. If it is PWV it is important to plot separately the PWV retrieved with SKYMAP in the comparisons sections. Are these retrievals useful or are just a step in the calculations to obtain the Fo?

Step 2 in SKYMAP algorithm retrieves the PWV, not the transmittance of the PWV. It is just a step in the calculations to obtain the F0. Instead of plotting SKYMAP-derived PWV, DSRAD-derived PWV with the calibration constant determined by the SKYMAP algorithm is compared in section 4.

Technical comments

- Abbreviations should be defined also in the abstract.

We agree with the reviewer. We revised it.

- L56-60. The two sentences should be separated. It seems that PWV is defined only at 940nm. 940 is a bandwidth that is selected because of the highest absorption in the shortwave spectral range.

115

We agree with the reviewer. We revised it (L79-84) as below:

Precipitable water vapor (PWV), which is the total atmospheric water vapor contained in a vertical column, has been estimated from the measurement of direct solar irradiance at the water vapor absorption bands. One of the strong water vapor absorption bands is around 940 nm and can be measured by sun photometer (Fowle, 1912, 1915; Bruegge *et al.*, 1992; Schmid *et al.*, 1996, 2001; Halthore *et al.*, 1997), SKYNET sky-radiometer (Campanelli *et al.*, 2014, 2018; Uchiyama *et al.*, 2014, 2018a), and AERONET sun-sky photometer (Holben *et al.*, 1998).

L66 Bruegge approach was also dependable on the altitude of the station, which made it difficult when transporting an instrument.

Yes, it is written in L.63-65 in the discussion paper and added the sentence in the revised manuscript (L87-90) as below:

However, there is a known noticeable uncertainty in the estimate of PWV because the adjustment parameters depend on the spectral sensitivity of the spectroradiometer as well as the vertical profiles of water vapor and temperature. Therefore, the adjustment parameters should be determined for each observation site.

Figure 10-11. It is not clear what is referred as true values. Please explain in the manuscript.

135

We agree with the reviewer. We revised it (L499-501) as below:

The retrievals of the volume size distribution, aerosol optical thickness, and PWV corresponded with their input values (“true” values in Fig. 11) when the input of PWV was <2 cm.

L124 It seems that something is missing. Which quantity is integrated from BOA to TOA?

We agree with the reviewer. We revised it (L162-165) as below:

In clear-sky conditions, the total optical thickness is the integrated value of aerosol scattering + absorption, Rayleigh scattering, and gas absorption coefficients in the column.

145

L152-153 How R was simulated? An RTM was used? Please describe in detail.

Yes, R is simulated by RTM (RSTAR). We revised it (L193-195) as below:

We examined the sensitivity of R at 940 nm in the two observation planes to PWV, aerosol optical properties, and aerosol vertical profiles by simulating R using the radiative transfer model RSTAR (Nakajima and Tanaka, 1986, 1988).

Figure 3. Please use same range in y-axis because it is confusing when it changes all the time.

We agree with the reviewer. We revised it as below:

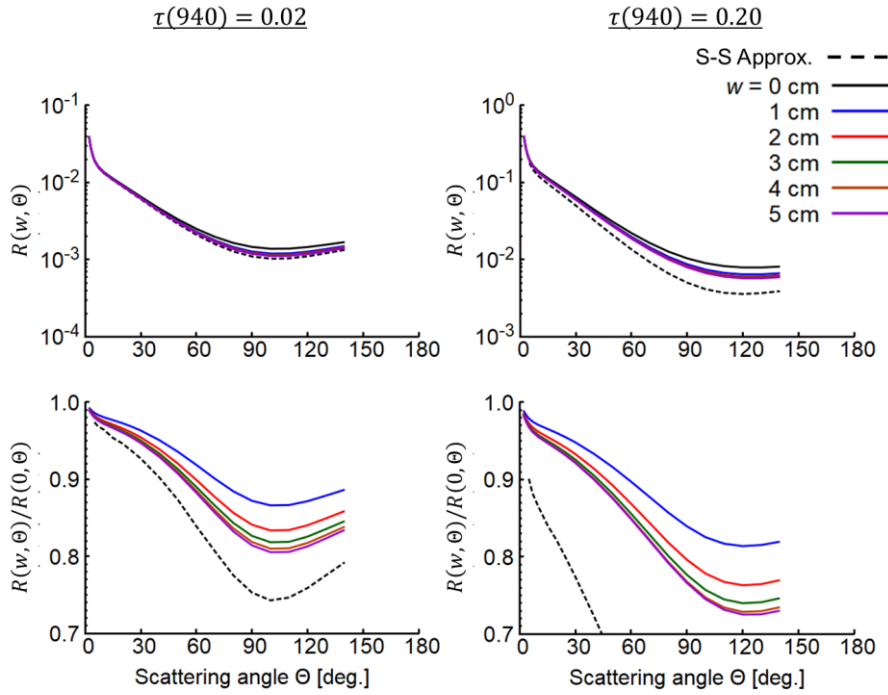


Figure 3: Normalized angular distributions simulated for continental average aerosol (Table 2) in the almucantar plane with aerosol optical thicknesses of 0.02 and 0.20 at 940 nm. Simulations were conducted for SZA = 70° and PWV (w) = 0, 1, 2, 3, 4, and 5 cm. The top row is the normalized radiance $R(w, \Theta)$, and the bottom row is the ratio of $R(w, \Theta)$ to $R(0, \Theta)$. S-S Approx. is single scattering approximation.

Some explanation should be provided regarding the selection of dust for the simulation.

We agree with the reviewer. We revised it (L216-218) as below:

The transported dust aerosol is composed of coarse particles, which have larger impacts on the angular distribution of R at the near-infrared wavelength than fine particles.

L240 There are not 18 boundary layers. As it stated by the term boundary, it is located on the boundaries. Stratosphere is not a boundary layer. I suggest to change to just “18 layers”/

The atmosphere consists 17 layers which boundary altitude is 18. Therefore, we revised it (L291-292) as below:

175 The model atmosphere is divided by 18 altitudes of 0, 1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 15, 20, 30, 40, 50, 70, and 120 km.

L440. 0.5cm is too big and not visible in figure 10. Since it is for values <2cm, this is more tha 50% error. Is it a typo or it is estimated from somewhere not shown on the figure?

180

We estimate the uncertainty from the sensitivity tests in Section 3. “0.5 cm” is a typo. We revised it (L522-524) as below:

When the input of PWV was < 2 cm, the SKYMAP algorithm retrieved PWV very well, within an error of 10% regardless of the aerosol optical thickness or aerosol type. This was also observed when the bias errors were added for *R*.

185

L476-85. Write in a clear manner that annual values refer to *F₀* and not PWV.

Yes, it refers to *F₀*. We wrote “the annual mean calibration constant” instead of “the annual mean value” in the revised manuscript.

190

Response to comments of referee 3

Authors would like to express sincere thanks to the referee 3 for valuable comments. We revised a manuscript carefully based on given comments. The comments of the referee 3 are in blue, our replies are in black, and changes made in the revised manuscript are in red. The English in this document has been checked by at least two professional editors, both native speakers of English. Our replies to the comments are as below.

This manuscript written by Momoi et al. presents a new in-situ method to calibrate the so-called water channel in a sky-radiometer Prede POM02, and to derive the precipitable water vapor content in the atmospheric column, by using measurements from the sky-radiometer in the almucantar and principal planes. The method means a further step in the development of methods for the retrieval of aerosols and gases, with very good applicability for the SKYNET international network.

The method description is also complemented with sensitivity analysis, ending with a couple of experimental cases along 1-2 years in two different sites.

The paper is clear, with a good detail about the theoretical considerations. The English grammar is good, with very few typo errors if any. I consider the manuscript must be accepted in AMT, after some comments are addressed.

General comments:

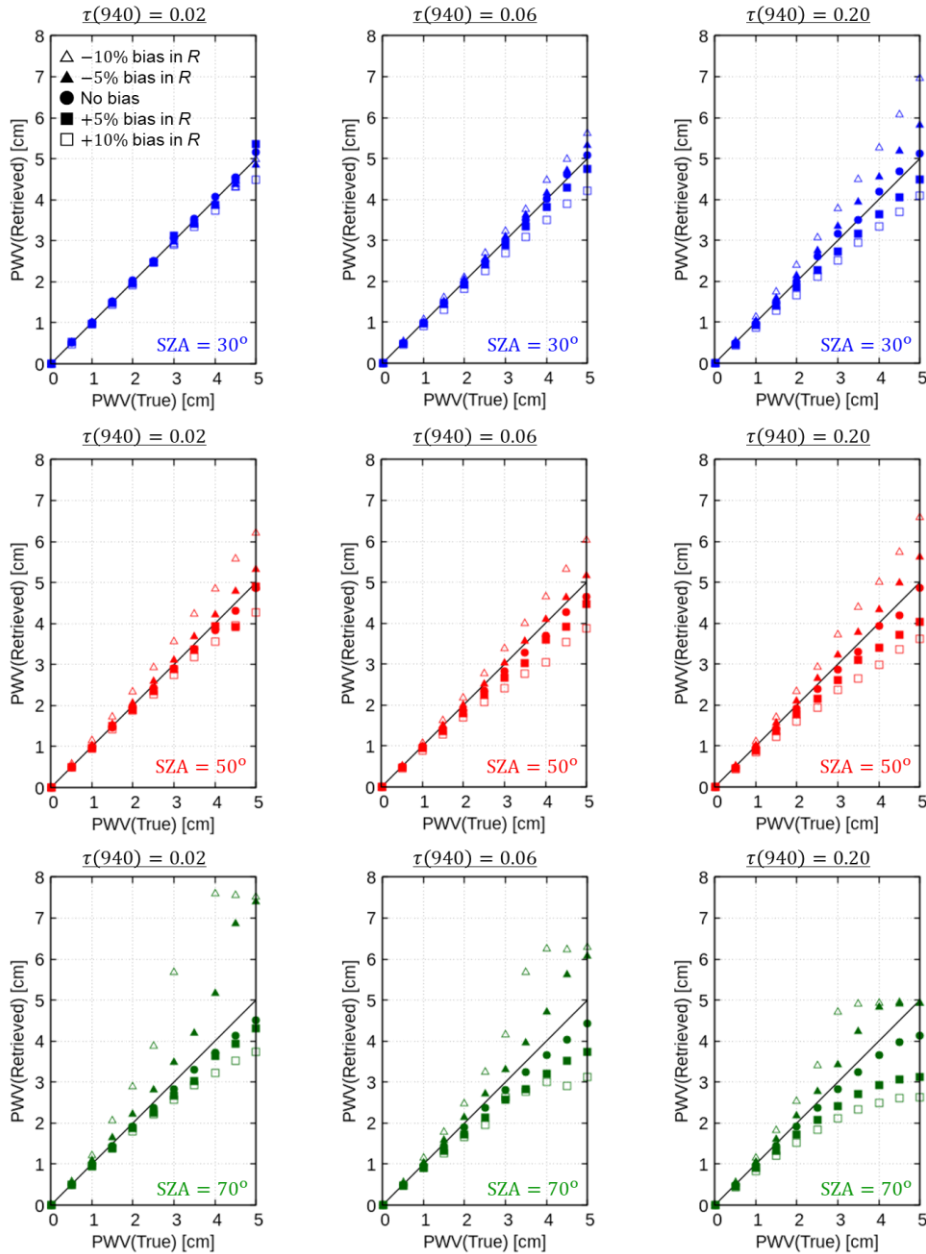
- It is very important to include an estimation of the uncertainty of the method, performed by expanding the forward sensitivity analysis or by comparison with other methods, even if preliminary. This information is critical and I miss it in the abstract and more in depth in the text body, if it is currently available.

- Errors in aerosol properties interpolated at 940 nm could have an impact on the comparison of PWV between AERONET and this method? Have the authors checked the comparison of aerosol properties from Cimel and this method, during the period studied? If yes, it is not need to include a detailed description, but comment (related to Figure 14).

We discussed the uncertainty of PWV by conducted the sensitivity tests using the simulated data with the bias errors in diffuse radiances in Section 3 and compared with the AERONET sun-sky radiometer in Section 4.2. We also estimate error propagation from aerosol optical thickness to PWV in Appendix B in the revised manuscript as below:

(Section 3: L510-521)

We also conducted sensitivity tests using the simulated data with bias errors to investigate uncertainty in the SKYMAP-derived PWV. The bias errors were $\pm 5\%$ and $\pm 10\%$ for R . The value of 5% was given by following reasons. The SVA bias errors of the diffuse radiances for the sky-radiometer observations were estimated to be less than 5% (Uchiyama *et al.*, 2018b). According to Dubovik *et al.* (2000), the uncertainty of the diffuse radiances for the AERONET measurements is $\pm 5\%$. Figures 13 and 14 show the results from the simulated data for the continental average and transported dust aerosols with aerosol optical thicknesses of 0.02, 0.06 and 0.20 at 940 nm. PWV was overestimated when -5% bias was applied to R . This corresponds to the relationship between R and PWV, where R decreases with increasing PWV (Section 2.1.2). The bias errors strongly affected the retrieval of PWV at high PWV (> 2 cm), because the sensitivity of high PWV is lower than that of low PWV. The retrieval error of PWV increased with increasing bias errors. The retrieval error of PWV due to $\pm 5\%$ and $\pm 10\%$ errors for R was within 10% for $\text{PWV} < 2$ cm and up to 200% for $\text{PWV} > 2$ cm.



50 **Figure 13: Comparison of the “true” and retrieval values of PWV from simulated data for continental average aerosol with bias errors. The top, middle, and bottom rows are the retrieval results at $SAZ = 30^\circ$, 50° , and 70° , respectively. Closed circles are the results with no bias errors. Closed squares and closed triangles are the results with bias errors of plus and minus 5% in R , respectively. Open squares and open triangles are the results with bias errors of plus and minus**
 55 **10% in R , respectively.**

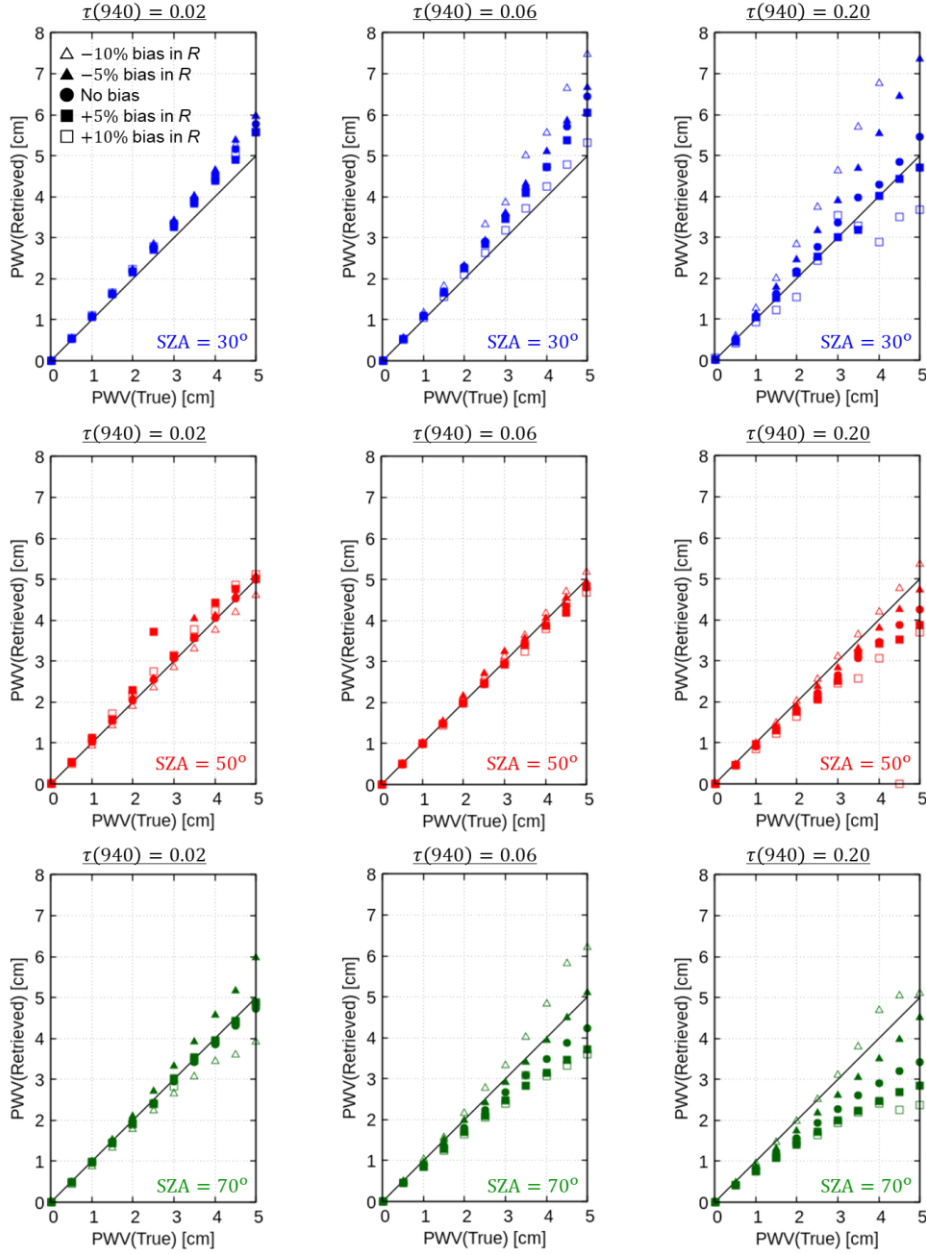


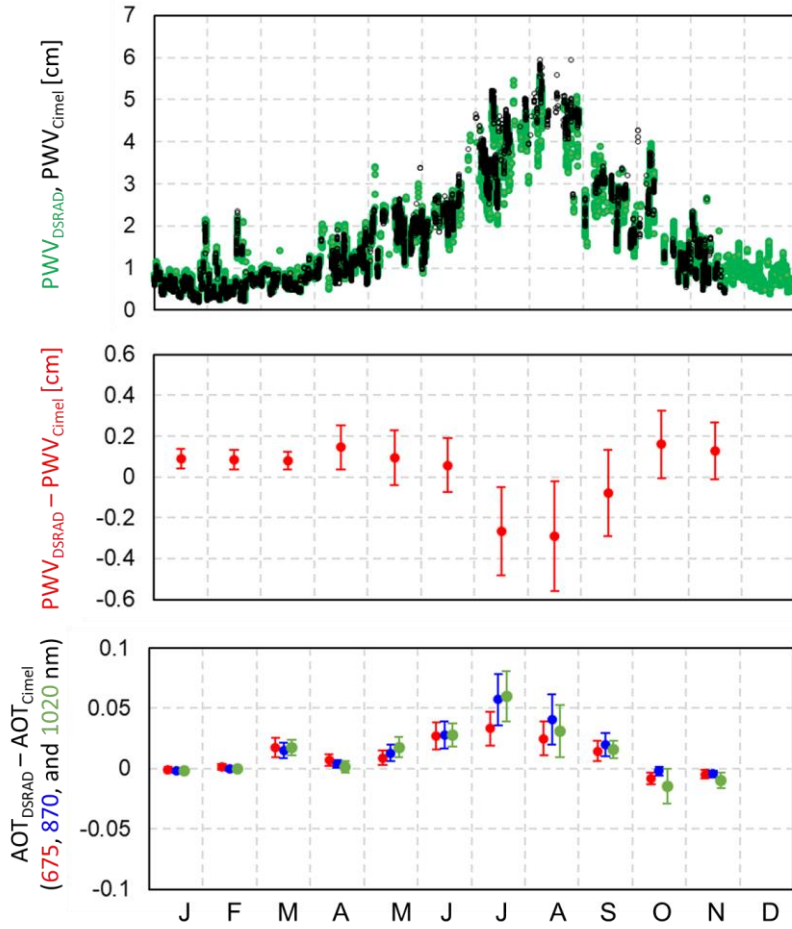
Figure 14: Similar to Fig. 13 but for transported dust aerosol.

60 (Section 4.2: L607-625)

PWV_{DSRAD+SKYMAP} using the annual mean calibration constant was 12% and 9.1% smaller than PWV_{MWR} and PWV_{Cimel}, respectively (Table 5). These results suggest an underestimation of PWV_{DSRAD+SKYMAP}, as the uncertainty of PWV_{Cimel} compared to the GNSS/GPS receiver is expected to

be less than 10% (Giles *et al.*, 2018). The underestimation of $PWV_{DSRAD+SKYMAP}$ was due to two factors.

65 The first is the retrieval of PWV by the annual mean calibration constant for the water vapor channel. The calibration constant not only is subject to aging but also undergoes seasonal variation due to temperature dependency (Uchiyama *et al.*, 2018a). Thus, it is possible to underestimate the calibration constant in the wet season. Second, uncertainty regarding the aerosol optical thickness affected PWV retrieval. Figure 18 depicts the differences in PWV and aerosol optical thicknesses at 675, 870, and 1020 nm between the DSRAD algorithm and the AERONET retrieval. In the periods from January to May and from October to November, the differences in PWV and aerosol optical thicknesses were less than 0.1 cm and 0.015, respectively. However, the difference in PWV was greater than 0.1 cm from July to September. This corresponds to the difference in aerosol optical thicknesses at 675, 870, and 1020 nm from July to September, which indicates that the transmittance of water vapor was overestimated by the overestimation of aerosol optical thickness. This led to the underestimation of $PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant when PWV was > 3 cm. In our error estimation, the error of $+ 0.03$ for the aerosol optical thickness at 940 nm resulted in the error of $- 0.214$ cm for PWV (Appendix B).



80

Figure 18: The top row shows the time series of PWV in 2017 at Chiba (green and black circles are $PWV_{DSRAD+SKYMAP}$ and PWV_{Cimet} , respectively). The middle row is the difference between

PWV_{DSRAD+SKYMAP} and PWV_{Cimel}. The bottom row is the difference in aerosol optical thicknesses at 675 nm (red), 870 nm (blue), and 1020 nm (green) between the DSRAD algorithm and the AERONET retrieval results. Circles and error bars in the middle and bottom rows are means and standard deviations, respectively.

Table 5: Comparison of PWV between DSRAD and other instruments.

		Slope C_1	Intercept C_2 [cm]	γ	Bias [cm]	RMSE [cm]
PS1202091 at Tsukuba, Japan						
Monthly mean F_0	vs GNSS/GPS receiver (2013)	0.956	0.079	0.938	-0.049	0.138
	vs GNSS/GPS receiver (2014)	0.937	0.161	0.970	-0.110	0.170
Annual mean F_0	vs GNSS/GPS receiver (2013)	0.919	0.173	0.987	-0.061	0.226
	vs GNSS/GPS receiver (2014)	0.934	0.178	0.987	-0.089	0.223
PS2501417 at Chiba, Japan						
Monthly mean F_0	vs MWR (2017)	0.964	0.053	0.961	-0.027	0.091
	vs AERONET (2017)	0.987	0.107	0.976	0.098	0.122
Annual mean F_0	vs MWR (2017)	0.880	0.132	0.985	0.042	0.231
	vs AERONET (2017)	0.909	0.184	0.991	0.055	0.186

C_1, C_2 : $\text{PWV}_{\text{DSRAD}} = C_1 \times \text{PWV}_{\text{Other}} + C_2$
Bias: $\text{PWV}_{\text{DSRAD}} - \text{PWV}_{\text{Other}}$

(Appendix B)

Appendix B: Error propagation from aerosol optical thickness to PWV

We evaluated the influence of the uncertainty of aerosol optical thickness on PWV using the empirical equation of Bruegge *et al.* (1992). PWV is described using the adjustment parameters as follows

$$w = \frac{1}{m_0} \left(-\frac{\ln \bar{T}_{\text{H}_2\text{O}}}{a} \right)^{\frac{1}{b}} [\text{cm}]. \quad (\text{B1})$$

The uncertainty of PWV ϵ_{PWV} is given from the partial differentiation of Eq. (B1) with respect to $\ln \bar{T}_{\text{H}_2\text{O}}$ as follows

$$\epsilon_{\text{PWV}} = \frac{\partial w}{\partial \ln \bar{T}_{\text{H}_2\text{O}}} \epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}} = \frac{w}{b \ln \bar{T}_{\text{H}_2\text{O}}} \epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}}. \quad (\text{B2})$$

where $\epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}}$ is the uncertainty of $\bar{T}_{\text{H}_2\text{O}}$. Using Eq. (B1) with the adjusting parameters of the sky-radiometer, with $a = 0.620$ and $b = 0.625$ as the coefficient values for the trapezoidal spectral response function (Uchiyama *et al.*, 2018a), we write the uncertainty of PWV as

$$\epsilon_{\text{PWV}} = -\frac{w}{ab} (m_0 w)^{-b} \epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}} = -\frac{w}{0.388} (m_0 w)^{-0.625} \epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}}. \quad (\text{B3})$$

If the uncertainty of the calibration constant at the water vapor channel is ignored, the uncertainty of $\bar{T}_{\text{H}_2\text{O}}$ is given from Eq. (21) as follows

$$\epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}} = m_0 \epsilon_{\text{AOT}}. \quad (\text{B4})$$

where ϵ_{AOT} is the uncertainty of the aerosol optical thickness at 940 nm. The uncertainty of PWV is written by Eqs. (B3) and (B4) as

$$\epsilon_{\text{PWV}} = -\frac{1}{0.388} (m_0 w)^{0.375} \epsilon_{\text{AOT}} = -0.214 [\text{cm}]. \quad (\text{B5})$$

where $m_0 = 3.0$, $w = 5.0$ cm, and $\epsilon_{\text{AOT}} = 0.03$.

120

- The POM02 has been used in this study; however, version POM01 also has a channel in 940nm. Then I understand the method can also be applied to this version. This should be made clear.

We agree with the reviewer. Our methods (SKYMAP and DSRAD) can be used for sky-radiometer model POM-01. Thus, Section 1 and 2 of the revised manuscript was generally written.

125

- What is the model used for the sensitivity analysis performed in section 2.1.2? RSTAR? State it clearly and give some detail.

130 Yes, R is simulated by RTM (RSTAR). We revised it (L193-195) as below:

We examined the sensitivity of R at 940 nm in the two observation planes to PWV, aerosol optical properties, and aerosol vertical profiles by simulating R using the radiative transfer model RSTAR (Nakajima and Tanaka, 1986, 1988).

135 - A comparison to other techniques for PWV retrieval developed in SKYNET framework is envisaged (i.e. Campanelli et al), and discussion of pro/cons advised. If this is planned for a future work, please state.

We agree with the reviewer. We revised it (L656-658) as below:

140 In future work, we plan to compare our method with others in the SKYNET framework (Uchiyama *et al.* 2014; Campanelli *et al.*, 2014).

- Is the cloud screening also a new method developed for this study? Two different thresholds are used for the two near and far indices. How these thresholds have been selected? Did you performed an statistically study, or it is simply a preliminary proposal?

Yes, the cloud screening method, called SCAD method, is developed for this study. We determined the thresholds by comparing the images of the whole-sky camera and the time series of the surface solar radiation observed by the pyranometer and validated by visual observation. We added the validation results of the SCAD method in the revised manuscript (L455-464) as below:

The results were validated using visual observation of the amount of clouds in the Aerological Observatory of the JMA. Figure 10a shows the histograms of index 1 for cases in which the sun was and was not covered by clouds. Index 1 had a low value when there were no clouds shading the sun but had a wide range of values when clouds were shading the sun. Fig. 10b shows the histograms of index 2 when cloud cover was and was not $< 20\%$. The peak shifted to the right when cloud cover was $\geq 20\%$, but the effect was not significant. Table 3 shows the validation results of this method. We defined “best condition” as cloud cover $< 20\%$ and “poor condition” as cloud cover $\geq 20\%$. In less than 17% of cases a “poor condition” was judged as a “best condition”. The sky-radiometer observes only a part of the whole sky, but our algorithm showed good results.

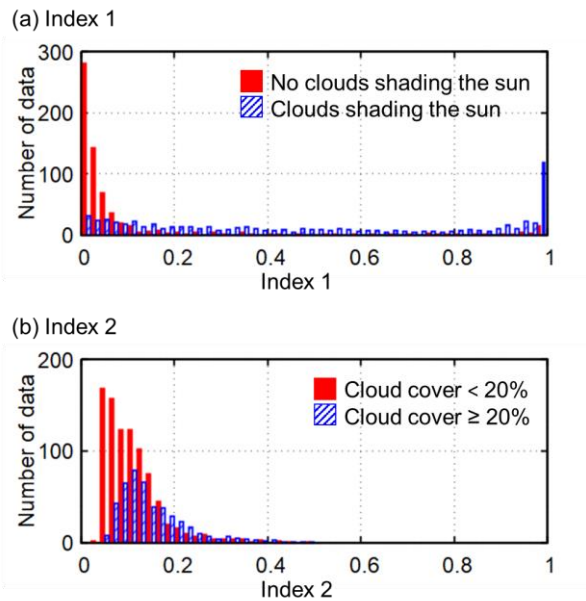


Figure 10: Histograms of indexes 1 and 2 of sky-radiometer observations at Tsukuba. (a) Index 1 when the sun is covered by clouds (blue boxes) and not covered by clouds (red boxes). (b) Index 2 when cloud cover is less than to 20% (red boxes) and greater than or equal to 20% (blue boxes).

Table 3: Validation of the SCAD method by visual observation from 2013 to 2014 in Tsukuba.

Visual observation Cloud cover	Sky-radiometer measuring plane	
	Available condition	Poor quality condition
Clear, less than 20%	463 (83.4%)*	68 (8.7%)
Cloud affected, more than 20%	92 (16.6%)	714(91.3%)*

*Obviously correct determination.

Technical comments

170

- Abstract, line 22: "remains challenging" ?

Yes, it is. We revised it (L28) as below:

...the sky-radiometer remains challenging...

175

- Abstract, line 28-30: in the second step, the transmittance of PWV is retrieved; however, both in the text and the plots look like it is the PWV content, not the transmittance, retrieved. Please write consistently.

180 The step 2 of the SKYMAP retrieves PWV to estimate the transmittance of the PWV, but not directly retrieves the PWV. We revised it (L39).

- page 3, line 73: the calibration method described for AERONET is solely for field instruments, please state.

185

We agree with the reviewer. We revised it (L97-100) as below:

In the AERONET led by NASA, the field instrument of the AERONET sun-sky radiometer is calibrated every year by lamp calibration and side-by-side comparison with a reference spectroradiometer (Holben *et al.*, 1998).

190

- page 6, line 150: I'm not sure to understand the need for this sentence here, or it looks like incomplete discussion here.

We agree with the reviewer. It is deleted and mentioned in Sect. 2.1.2 in the revised manuscript.

195

- Page 11, line 326: In equation 21, the band average transmittance is used, converted from PWV in step 2. Do you mean the conversion is performed in the step 3, from the PWV obtained in step 1?

The band average transmittance in Eq. 21 is converted from the PWV retrieved in step 2 not step 1.
200 Since the diffuse sky irradiances are calculated for each sub-band by the radiative transfer model, in step 2 you can get the PWV instead of the band average transmittance of the PWV and need to convert it to the band average transmittance to get the F0.

- Page 12, line 355. Is the Huber's M-estimation method iterative? (the weight to calculate $\ln\langle F0 \rangle$ depends on $\ln\langle F0 \rangle$, so any condition applied?)
205

Yes, Huber's M-estimation method is iterative. This method has the potential to apply noisy satisfying normal distribution, not white noise.

210 - Page 14, line 415: what is the time stamp used for any triplet? The central element time stamp or the first element time stamp?

The time stamp used the central element time stamp. We revised it (L485-486) as below:

Second, the triplet of the aerosol optical thickness in Smirnov *et al.* (2000) is built from the pre/post 1
215 min data instead of 30 s.

- Page 14, line 440: this error estimation about 0.5cm looks somewhat high to me, in comparison to other techniques. It would be useful to provide literature estimations of the uncertainty for the methods used as referents (AERONET, MWR, GPS...) in the comparison discussion.
220

We estimate the uncertainty from the sensitivity tests in Section 3. "0.5 cm" is a typo. We compared our retrievals with more establish methods, such as GNSS/GPS-derived PWV, and AERONET-derived PWV as below:

(Section 4.1: L582-587)

225 Table 5 summarizes the results of comparisons of DSRAD-derived PWV and GNSS/GPS-derived PWV. The magnitude of the bias error and root mean square error were small, less than 0.11 cm and less than 0.226 cm, during 2013 to 2014. Table 6 shows the errors of the retrieved PWV with the annual mean calibration constants for the rank of PWV. The bias error was larger for high PWV than it was for low PWV. The magnitude of the bias errors of PWV was less than 0.163 cm for $PWV < 3$ cm and less than
230 0.339 cm for $PWV > 3$ cm.

(Section 4.2: L602-610)

$PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant agreed with PWV_{MWR} (Fig. 17c). The error of $PWV_{DSRAD+SKYMAP}$ was $-0.041 < \text{bias} < 0.024$ cm and $RMSE < 0.212$ cm for low PWV (< 3

cm) and bias < -0.356 cm and RMSE > 0.465 cm for high PWV (Table 6). Figure 17d shows that PWV_{DSRAD+SKYMAP} using the annual mean calibration constant also agreed with PWV_{Cimel} for low PWV (< 3 cm) but was smaller than PWV_{Cimel} for high PWV (> 3 cm). PWV_{MWR} was larger than PWV_{Cimel} (Fig. 17e). PWV_{DSRAD+SKYMAP} using the annual mean calibration constant was 12% and 9.1% smaller than PWV_{MWR} and PWV_{Cimel}, respectively (Table 5).

240

Table 5: Comparison of PWV between DSRAD and other instruments.

		Slope C_1	Intercept C_2 [cm]	γ	Bias [cm]	RMSE [cm]
PS1202091 at Tsukuba, Japan						
Monthly mean F_0	vs GNSS/GPS receiver (2013)	0.956	0.079	0.938	-0.049	0.138
	vs GNSS/GPS receiver (2014)	0.937	0.161	0.970	-0.110	0.170
Annual mean F_0	vs GNSS/GPS receiver (2013)	0.919	0.173	0.987	-0.061	0.226
	vs GNSS/GPS receiver (2014)	0.934	0.178	0.987	-0.089	0.223
PS2501417 at Chiba, Japan						
Monthly mean F_0	vs MWR (2017)	0.964	0.053	0.961	-0.027	0.091
	vs AERONET (2017)	0.987	0.107	0.976	0.098	0.122
Annual mean F_0	vs MWR (2017)	0.880	0.132	0.985	0.042	0.231
	vs AERONET (2017)	0.909	0.184	0.991	0.055	0.186

C_1, C_2 : $PWV_{DSRAD} = C_1 \times PWV_{Other} + C_2$
Bias: $PWV_{DSRAD} - PWV_{Other}$

245 **Table 6: Difference in PWV between DSRAD with the annual mean calibration constants and other instruments.**

	PWV _{Other}				
	0 – 1 cm	1 – 2 cm	2 – 3 cm	3 – 4 cm	> 4 cm
	Bias [cm] (RMSE [cm])	Bias [cm] (RMSE [cm])	Bias [cm] (RMSE [cm])	Bias [cm] (RMSE [cm])	Bias [cm] (RMSE [cm])
PS1202091 at Tsukuba, Japan					
vs GNSS/GPS receiver (2013)	0.083 (0.124)	0.160 (0.211)	0.084 (0.236)	-0.098 (0.326)	-0.339 (0.537)
vs GNSS/GPS receiver (2014)	0.110 (0.142)	0.163 (0.221)	0.107 (0.251)	-0.055 (0.353)	-0.239 (0.492)
PS2501417 at Chiba, Japan					
vs MWR (2017)	0.017 (0.066)	0.024 (0.153)	-0.041 (0.212)	-0.356 (0.465)	-0.594 (0.722)
vs AERONET (2017)	0.088 (0.105)	0.118 (0.192)	0.017 (0.223)	-0.214 (0.386)	-0.264 (0.306)

Bias: $PWV_{DSRAD} - PWV_{Other}$

250 - Page 17, summary (conclusions): is there any error estimation for PWV >2 cm? Some discussion on the pros and cons of this method and other methods developed for Prede POM instruments would be useful.

We added the results of the uncertainty for PWV >2 cm in the revised manuscript (L650-655) as below:
The magnitude of the bias error and the root mean square error were < 0.163 cm and < 0.251 cm, respectively, for low PWV (< 3 cm). However, our retrieved PWV was underestimated in the wet

255 conditions, and the magnitude of the bias error and the root mean square error were less than 0.594 cm
and less than 0.722 cm for high PWV. This was due to seasonal variation in the calibration constant and
the overestimation of aerosol optical thickness at 940 nm interpolated from those at 870 and 1020 nm.

- Is figure 1 used in the text?

260 It is using the top of Section 2 and Section 3 in the discussion paper.

- Caption figure 3: please explain what S-S refers to (I think it is single scattering) .

265 Yes, “S-S approx.” means the single scattering approximation. We added the sentence in the caption as
below:

S-S Approx. is single scattering approximation.

- Figures: I think saying "top row" and "bottom row" is more appropriate than "top and bottom line".

270 We agree with the reviewer. We revised it.

Development of on-site self-calibration and retrieval methods for sky-radiometer observations of precipitable water vapor

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Abstract.

The Prede sky-radiometer measures direct solar irradiance and the angular distribution of diffuse
20 radiances at the ultraviolet, visible, and near-infrared wavelengths. These data are utilized for remote
sensing of aerosols, water vapor, ozone, and clouds, but the calibration constant which is the sensor
output current of the extraterrestrial solar irradiance at the mean distance between the Earth and the sun,
is needed., whose aerosol channels are calibrated by on-site measurements (the Improved Langley
25 method), has The aerosol channels, which are the weak gas absorption wavelengths of 340, 380, 400,
500, 675, 870, and 1020 nm, can be calibrated by an on-site self-calibration method, the Improved
Langley method. This on-site self-calibration method is ~~useful~~been used for the continuous long-term
observation of aerosol properties. However, the continuous long-term observation of precipitable water
vapor (PWV) by the sky-radiometer remains challenging, because ~~the-calibrating the~~ water vapor
~~channel~~absorption channel of 940 nm is generally ~~relies~~calibrated by on the standard Langley method
30 (SL) at limited observation sites (e.g., the Mauna Loa Observatory) and the transfer of the calibration
constant by side-by-side comparison with the reference sky-radiometer calibrated by the SL method. In
this study, we developed the SKYMAP algorithm, a new on-site ~~self-calibration~~ method ~~for-of self-~~
~~calibrating~~ the water vapor channel of the ~~Prede~~ sky-radiometer using diffuse radiances normalized by
direct solar irradiance (normalized radiances). Because the sky-radiometer measures direct solar
35 irradiance and diffuse radiance using the same sensor, the normalization cancels the calibration constant
included in the measurements. The SKYMAP algorithm consists of three steps. First, aerosol optical
and microphysical properties are retrieved using direct solar irradiances and ~~the-normalized-diffuse~~
radiances at aerosol channels. The aerosol optical properties at the water vapor channel are interpolated
from those at aerosol channels. Second, ~~the transmittance of~~ PWV is retrieved using ~~the diffuse radiance~~
40 ~~normalized to the direct solar irradiance~~the angular distribution of the normalized radiances at the water
vapor channel, ~~which does not need the calibration constant~~. Third, the calibration constant at the water
vapor channel is estimated from the transmittance of PWV and aerosol optical properties. Intensive
sensitivity tests of the SKYMAP algorithm using simulated data of the sky-radiometer showed that the
calibration constant is retrieved reasonably well for $PWV < 2$ cm, ~~indicating which indicates~~ that the
45 SKYMAP algorithm can calibrate the water vapor channel on-site in dry conditions. ~~Then-Next, the~~
SKYMAP algorithm was applied to actual measurements under the clear-sky and low PWV (< 2 cm)
conditions in the dry season at two sites, (Tsukuba and Chiba, Japan, and the annual mean calibration
constants at the two sites were determined.). ~~Because the SKYMAP algorithm is useful for clear-sky~~
~~and low PWV (< 2 cm) conditions, the water vapor channel was calibrated for the dry season. The~~
50 SKYMAP-derived calibration constants were 10.1% and 3.2% lower, respectively, than those
determined by side-by-side comparison with the reference sky-radiometer. After determining the
calibration constant, we obtained ~~PWV is able to be retrieved using from the~~ direct solar irradiances ~~for~~
~~the whole year in both the dry and wet seasons~~. The retrieved PWV values corresponded well to those

derived from a Global Navigation Satellite System-~~(GNSS)~~/Global Positioning System ~~(GPS)~~-receiver,
55 a microwave radiometer, and an AERONET sun-sky radiometer at both sites. ~~The~~ (correlation
coefficients ~~were greater than~~ $\gamma > 0.96$. ~~We calculated the bias errors and the root mean square errors by~~
~~comparing PWV between the DSRAD algorithm and other instruments. The magnitude of the bias error~~
~~and the root mean square error were < 0.163 cm and < 0.251 cm for $PWV < 3$ cm, respectively.~~
60 ~~However, our method tended to underestimate PWV in the wet conditions, and the magnitude of the~~
~~bias error and the root mean square error became large, < 0.594 cm and < 0.722 cm for $PWV > 3$ cm,~~
~~respectively.). This problem was mainly due to the overestimation of the aerosol optical thickness~~
~~before the retrieval of PWV. These results show~~~~indicating~~ that the ~~Prede-sky-radiometer provides both~~
~~aerosol and~~SKYMAP algorithm enables us to observe PWV ~~over the long term,~~data based on its unique
on-site self-calibration methods.

65

1 Introduction

The highly variable spatiotemporal distributions of aerosols, clouds, and gases (*e.g.*, water vapor and ozone) still include large uncertainties for the quantitative understanding of the Earth's radiation budget at various spatial and temporal scales. Water vapor is specified as an essential climate variable (ECV) by the World Meteorological Organization (WMO), a critical key parameter that contributes to characterizing Earth's climate and changes in atmospheric temperature (Schmidt *et al.*, 2010). Water vapor absorbs visible radiation and absorbs and emits infrared radiation to heat and cool the Earth and its atmosphere. Atmospheric heating drives the evaporation of sea water, causing an increase in temperature as positive feedback (IPCC, 2013). In addition, the distribution of water vapor controls precipitation amounts and aerosol-cloud interactions (Twomey, 1990). To understand these effects quantitatively, many previous studies have measured ~~precipitable columns of~~ water vapor ~~content by~~ using a radiosonde, Global Navigation Satellite System (GNSS)/Global Positioning System (GPS) receiver (Bevis *et al.*, 1992), or spectroradiometer (*e.g.*, Fowle, 1912, 1915).

Precipitable water vapor (PWV), which is the total atmospheric water vapor contained in a vertical column, has been estimated from the measurement of direct solar irradiance at the water vapor ~~channel-absorption bands. One of the strong water vapor absorption bands is at~~ around 940 nm ~~and can be measured~~ by sun photometer (Fowle, 1912, 1915; Bruegge *et al.*, 1992; Schmid *et al.*, 1996, 2001; Halthore *et al.*, 1997), SKYNET sky-radiometer (Campanelli *et al.*, 2014, 2018; Uchiyama *et al.*, 2014, 2018a), and AERONET sun-sky photometer (Holben *et al.*, 1998). ~~These previous studies of SKYNET and AERONET~~ derived PWV from the observed transmittance of water vapor ($\bar{T}_{\text{H}_2\text{O}}$), assuming $\bar{T}_{\text{H}_2\text{O}} = e^{-a(m \cdot w)^b}$ (Bruegge *et al.*, 1992), where a and b are ~~adjustment adjusting~~ parameters, m is the optical air mass, and w is PWV. However, there is a known noticeable uncertainty in the estimate of PWV because the adjustment parameters depend on ~~radiometric calibration~~ the spectral sensitivity of the spectroradiometer as well as the vertical profiles of water vapor and temperature. ~~Therefore, the adjustment parameters should be determined for each observation site. Therefore,~~ Campanelli *et al.* (2014, 2018) developed a practical method for determining the ~~adjustment adjusting~~ parameters based on PWV retrieved by a GNSS/GPS receiver or by surface humidity observations.

To estimate PWV using a spectroradiometer, it is necessary to calibrate the water vapor channel. The calibration constant, ~~which is the sensor output current of the extraterrestrial solar irradiance at the mean distance between the Earth and the sun,~~ at the water vapor channel can be determined by the Langley method. For example, Uchiyama *et al.* (2014) calibrated the water vapor channel of a sky-radiometer with high accuracy using observations from the Mauna Loa Observatory (3400 m a.s.l.). In the AERONET led by NASA, the field instrument of the AERONET sun-sky radiometer is calibrated every year by lamp calibration and side-by-side comparison with a reference spectroradiometer (Holben

100 *et al.*, 1998). Dedicated effort and ~~cost expenses~~ are required ~~for maintaining to maintain~~ accurate long-term calibrations using these methods.

The ~~S~~sky-radiometer ~~models~~ POM-01 and POM-02 (Prede, Tokyo, Japan), which ~~is~~ are deployed in the international radiation observation network SKYNET, measures solar direct irradiances and ~~angular distributions of~~ diffuse irradiances at ~~the ultraviolet, visible, and near-infrared wavelengths.~~ 11 ~~wavelengths including~~ These measurements are used for the remote sensing of aerosol, cloud, water vapor, and ozone ~~channels~~ (Table 1; Takamura and Nakajima, 2004; Nakajima *et al.*, 2007). ~~Table 1 shows the relationship between the wavelengths and the main target of the remote sensing. The aerosol channels are 340, 380, 400, 500, 675, 870, and 1020 nm; the water vapor channel is 940 nm; the ozone channel is 315 nm; and the cloud channels are 1225, 1627, and 2200 nm. Through~~ 110 ~~on-site self-calibration of the aerosol channels by the Improved Langley (IL) method (Tanaka *et al.*, 1986; Nakajima *et al.*, 1996; Campanelli *et al.*, 2004, 2007).~~ The SKYNET system is capable of long-term and ~~automatic continuous~~ aerosol observation. ~~through an on-site self-calibration method, called the Improved Langley (The IL) method, which works not only in clean atmospheric conditions, but also in turbid atmospheric conditions (Tanaka *et al.*, 1986; Nakajima *et al.*, 1996; Campanelli *et al.*, 2004, 2007).~~ 115 However, no improved calibration method has replaced the standard (Uchiyama *et al.*, 2014) or modified (Campanelli *et al.*, 2014, 2018) Langley methods for the water vapor channel. In this study, we ~~devised developed~~ a new method of retrieving PWV using the PWV dependency of the normalized radiance, defined as the ratio of diffuse radiance to direct solar irradiance at the water vapor channel. This method enables us to estimate PWV without the calibration constant, and to perform on-site self-calibration of the water vapor channel. We developed two algorithms, SKYMAP and DSRAD. 120 The SKYMAP algorithm is a new on-site method for self-calibrating the water vapor channel. It retrieves PWV (PWV_{SKYMAP}) from the angular distribution of the normalized radiance at ~~940 nm the~~ water vapor channel and calibrates the water vapor channel. The DSRAD algorithm estimates ~~the~~ PWV (PWV_{DSRAD}) from the calibrated direct solar irradiance at ~~940 nm the~~ water vapor channel. This method 125 does not require adjustment parameters and explicitly uses the filter response function and the vertical profiles of water vapor, temperature, and pressure. The SKYMAP and DSRAD algorithms are described in Section 2. We discuss the results of sensitivity tests of the ~~two~~ SKYMAP algorithms using simulation data in Section 3 and apply ~~the two~~ algorithms to observational data at two SKYNET sites in Section 4. At these two sites, PWV is observed by the GNSS/GPS receiver, MWR, or AERONET sun- 130 sky radiometer other than the sky-radiometer. The retrieval accuracy of our method is evaluated by comparison to these established methods.

2 Methods

In this study, PWV is retrieved using angular distributions of the normalized radiance, which does not require the calibration constant of the sky-radiometer. Section 2.1 shows the normalized radiances and dependencies of the normalized radiance on PWV. Next, we describe two algorithms, the flow and relationships of which are shown in Fig. 1. The SKYMAP algorithm retrieves aerosol optical and microphysical properties and calibrates the water vapor channel by retrieving PWV from the angular distribution of the normalized radiance (Section 2.2). The DSRAD algorithm retrieves PWV from the transmittance derived from the direct solar irradiance at the water vapor channel (Section 2.3).

2.1 Sky-radiometer measurements and the relationship between normalized radiances and the PWV

We explain the normalized radiance (Nakajima et al., 1996) in Section 2.1.1 and the theoretical relationship between the normalized radiances and PWV in Section 2.1.2.

2.1.1 Sky-radiometer measurements

The direct solar irradiance (F) and angular distribution of the diffuse ~~radiance~~ irradiance (L_V) are measured at seven wavelengths by the model POM-01 or 11-eleven wavelengths by the model POM-02, including aerosol ($\lambda = 340, 380, 400, 500, 675, 870, \text{ and } 1020 \text{ nm}$), cloud (1627 and 2200 nm), water vapor (940 nm), and ozone (315 nm) channels (Table 1). L_V is measured in the almucantar and principal planes (Fig. 2). The angular distribution of L_V is measured at scattering angles $\Theta = 2^\circ, 3^\circ, 4^\circ, 5^\circ, 7^\circ, 10^\circ, 15^\circ, 20^\circ, 25^\circ, 30^\circ, 40^\circ, 50^\circ, 60^\circ, 70^\circ, 80^\circ, 90^\circ, 100^\circ, 110^\circ, 120^\circ, 130^\circ, 140^\circ, 150^\circ, \text{ and } 160^\circ$ in the almucantar and principal planes, every 10 min. ~~Aerosol~~ The aerosol channels are calibrated with the IL method using the normalized radiance at $\Theta < 30^\circ$. F and $L_V(\Theta \geq 4^\circ)$ at the aerosol and water vapor channels are used in this study.

In the plane-parallel non-refractive atmosphere, ~~F~~ The direct solar irradiance at the bottom of the atmosphere (BOA) at the solar zenith angle (SZA) θ_0 and the solar azimuth angle ϕ_0 is derived from

$$F(\lambda) = \frac{F_0}{d^2} \exp(-m_0 \tau(\lambda)), \quad (1),$$

where F_0 is the calibration constant, ~~which is the sensor output current of the direct solar irradiance at the top of the atmosphere (TOA) when the distance between Earth and the sun is 1 AU;~~ d is the distance between Earth and the sun (AU); λ is the wavelength; τ is the total optical thickness; and m_0 is optical air mass, represented as $m_0 = 1/\cos\theta_0$ ~~in the plane-parallel nonrefractive atmosphere~~. In clear-sky conditions, the total optical thickness is the integrated value of ~~consists of~~ aerosol scattering + /absorption, Rayleigh scattering, and gas absorption coefficients and integrated from BOA to TOA in the

165 column. Assuming a narrow spectral band filter response function, the normalized radiance (R), which is the ratio of L_V to F at the zenith angle (θ) and the azimuth angle (ϕ), is obtained from the radiative transfer equation:

$$R(\Theta, \lambda) = \frac{L_V(\Theta, \lambda)}{F(\lambda)m_0\Delta\Omega} = \int_0^{\tau(\lambda)} \exp\left[(\tau - \tau')\left(\frac{1}{\mu_0} - \frac{1}{\mu}\right)\right] \omega'(\lambda, \tau')P'(\Theta, \lambda, \tau')d\tau' + Q(\Theta, \lambda) \quad (2)$$

170

where $P'(\Theta, \lambda, \tau')$ and $\omega'(\lambda, \tau')$ are, the total phase function and the total single scattering albedo, respectively, at the altitude $\tau = \tau'$, $\Delta\Omega$ is the solid view angle (or field of view); Q is the multiple scattering contribution; and

$$\begin{aligned} \cos \Theta &= \cos \theta \cos \theta_0 + \sin \theta \sin \theta_0 \cos(\phi - \phi_0) \quad (3) \\ \mu &= \cos \theta ; \mu_0 = \cos \theta_0 \end{aligned}$$

Noted that F_0 is cancelled by the normalization. In the second term of Eq. (2), the solid view angle of each wavelength can be retrieved from the angular distribution around the solar disk (Nakajima *et al.*, 1996; Boi *et al.*, 1999; Uchiyama *et al.*, 2018b). Eq. (2) can be simplified in the almucantar plane due to $\theta = \theta_0$:

$$R(\Theta, \lambda) = \int_0^{\tau(\lambda)} \omega'(\lambda, \tau')P'(\Theta, \lambda, \tau')d\tau' + Q(\Theta, \lambda) = \omega(\lambda)\tau(\lambda)P(\Theta, \lambda) + Q(\Theta, \lambda) \quad (4)$$

185 where $P(\Theta, \lambda)$ and ω are the total phase function and the total single scattering albedo, respectively. In contrast, normalized radiances in the principal plane can be described simply, similar to Eq. (4), if we assume that the atmosphere is a single layer:

$$R(\Theta, \lambda) = \frac{\mu_0^2}{\mu_0 - \mu} \omega(\lambda)P(\Theta, \lambda) \left[1 - \exp\left(\frac{\tau(\lambda)}{\mu_0} - \frac{\tau(\lambda)}{\mu}\right)\right] + Q(\Theta, \lambda) \quad (5)$$

190

~~Noted that real atmosphere is not a single layer (Torres *et al.*, 2014).~~

2.1.2 The relationship between normalized radiances at the water vapor channel and PWV

We examined the sensitivity of R at 940 nm in the two observation planes to PWV, aerosol optical properties, and aerosol vertical profiles by simulating R using the radiative transfer model RSTAR (Nakajima and Tanaka, 1986, 1988). The simulation was conducted with two aerosol types based on those used by Kudo *et al.* (2016): the continental average, and the continental average + transported dust in the upper atmosphere (Table 2). The continental average consisted of water-soluble particles,

soot particles, and insoluble particles (Hess *et al.*, 1999). Transported dust was defined as the mineral-transported component from Hess *et al.* (1999). Figure 3 shows the dependencies of R in the almucantar plane on PWV for continental average aerosol (Table 1) with aerosol optical thicknesses of 0.02 and 0.20 at 940 nm. The simulations were conducted for at an SZA of 70° in the almucantar plane. R decreases with increasing PWV regardless of the aerosol optical thickness, and the aerosol optical thickness does not affect this relationship. This suggests that PWV can be estimated from the normalized angular distribution, which is the angular distribution of R , without the calibration constant.

The dependencies of R on PWV cannot be observed in the radiative transfer using single scattering approximation in the almucantar plane. The first term of Eq. (4) is the normalized single scattering contribution and includes only the influences of aerosol and Rayleigh scattering. Note that this is true only for R , and not for L_V , because total optical thickness contributes to the single scattering approximation of L_V . However, the second term for the multiple scattering includes the influence of water vapor absorption and creates the dependencies of R on PWV. Figure 3 shows that the dependency of R on PWV at the forward scattering angles is not strong, but R at the backward scattering angles between 90° and 120° changes drastically with PWV. The range of the scattering angle for R is an important factor.

Figure 4 illustrates the dependency of R on PWV for different observation planes. The simulation was conducted for transported dust aerosol (Table 2) with an aerosol optical thickness of 0.06 at 940 nm at an SZA of 70° in the almucantar and principal planes. The transported dust aerosol is composed of coarse particles, which have larger impacts on the angular distribution of R at the near-infrared wavelength than fine particles. The dependency of R in the almucantar plane on PWV is the same as in Fig. 3. The dependency of R on PWV was also found in both observation the principal planes. R increases with increasing PWV at $\theta \ll \theta_0$ and decreases with increasing PWV at $\theta \gg \theta_0$. Although the dependency of R on PWV in the almucantar plane is strong at the backward scattering angles, that in the principal plane is strong at scattering angles between 60° and 90°. R in the principal plane is more sensitive to PWV than R in the almucantar plane because the normalized single scattering contribution in Eq. (5) includes not only Rayleigh and aerosol scattering but also gas absorption.

In theory, the maximum scattering angle of the principal plane is $\theta_0 + 90^\circ$ and that of the almucantar plane is $2\theta_0$. When the SZA is small, the principal plane has a broader scattering angle range than the almucantar plane. Therefore, the principal plane is more advantageous for PWV retrieval. Figure 5 is the same as Fig. 4 but for an SZA of 30°. Because the maximum scattering angle of the principal plane is obviously larger than that of the almucantar plane, PWV retrieval using the principal plane is more effective compared to that using the almucantar plane.

R in the principal plane is affected by the aerosol vertical profile, but this influence can be ignored for R in the almucantar plane (Torres *et al.*, 2014). Figure 6 shows the normalized angular distribution in the two observation planes for the different heights of the transported dust layer. It is

obvious that the normalized angular distribution in the principal plane is sensitive to the aerosol vertical
 235 profile. Consequently, the principal plane is useful for retrieving PWV when the aerosol vertical profile
 is known, but the almucantar plane is better when the aerosol vertical profile is not known. In this study,
 we used the normalized angular distribution in the almucantar plane because the aerosol vertical profile
 was not known. The influence of SZA on the retrieval of PWV is examined in Section 3.

2.2 SKYMAP algorithm

240 The SKYMAP algorithm consists of three steps (Fig. 7). First, aerosol optical and microphysical
 properties are retrieved from F and normalized angular distributions at aerosol channels. Second,
 aerosol optical properties at the water vapor channel are interpolated from those at aerosol channels.
 PWV is retrieved from the normalized angular distribution at the water vapor channel. Third, the
 calibration constant at the water vapor channel is estimated from PWV and the aerosol optical
 245 properties.

2.2.1 Step 1: Retrieval of aerosol optical and microphysical properties

Aerosol optical and microphysical properties are estimated from sky-radiometer measurements at
 aerosol channels using normalized angular distributions and transmittance ~~of the total extinction~~ $T =$
 $\frac{I_{\text{aer}} F d^2}{F_0}$ with an optimal estimation method similar to the AERONET and SKYNET retrievals (Dubovik
 250 and King, 2000; Dubovik *et al.*, 2006; Kobayashi *et al.*, 2006; Hashimoto *et al.*, 2012; Kudo *et al.*,
 2016). Estimated optical and microphysical properties are the real and imaginary parts of the refractive
 index at aerosol channels (~~340, 380, 400, 500, 675, 870, 1020 nm~~ Table 1), the volume size distribution,
 and the volume ratio of non-spherical particles to total particles in coarse mode. Hereafter, these are
 referred to as aerosol parameters.

255 In step 1, we construct the forward model to calculate the sky-radiometer measurements from
 the aerosol parameters. We assume that the aerosol volume size distribution in the radius range from
 0.02 to 20.0 μm consists of 20-modal lognormal ~~volume-size spectra distributions~~ as illustrated in Fig.
 8:

$$260 \quad \frac{dV(r)}{d\ln r} = \sum_{i=1}^{20} C_i \exp \left[-\frac{1}{2} \left(\frac{\ln r - \ln r_i}{s} \right)^2 \right], \quad (6)$$

$$\ln r_i = \ln(0.02\mu\text{m}) + \frac{2i-1}{2} \ln \Delta r, \quad (7)$$

$$s \equiv \frac{\ln \Delta r}{1.65\eta}, \quad (8)$$

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$$\ln \Delta r \equiv \frac{1}{20} (\ln(20\mu\text{m}) - \ln(0.02\mu\text{m})) = \frac{3}{20} \ln 10, \quad (9)$$

where C_i , r_i , and s are the volume, radius, and width of each lognormal function, respectively. η is the parameter to determine the width and is given by a fixed value (Appendix A). We can separate the size distribution into fine and coarse modes by giving the boundary radius r_b , which is obtained as the local minimum. Furthermore, we separate coarse mode into spherical and non-spherical particles:

$$\frac{dV(r)}{d\ln r} = \frac{dV_f(r)}{d\ln r} + (1 - \delta) \frac{dV_c(r)}{d\ln r} + \delta \frac{dV_c(r)}{d\ln r}, \quad (10)$$

where $\frac{dV_f(r)}{d\ln r}$ is fine mode, $\frac{dV_c(r)}{d\ln r}$ is coarse mode, and δ is the fraction of the non-spherical particles in coarse mode (Fig. 8). The aerosol optical properties are calculated from the size distribution and refractive index, similar to the methods of Kudo *et al.* (2016) and Dubovik *et al.* (2006), as follows:

$$\tau_{\text{ext/sca}}(\lambda) = \sum_k \frac{dV_f(r_k)}{d\ln r} K_{\text{ext/sca}}^S(\lambda, n, k, r_k) + \sum_k (1 - \delta) \frac{dV_c(r_k)}{d\ln r} K_{\text{ext/sca}}^S(\lambda, n, k, r_k) + \sum_k \delta \frac{dV_c(r_k)}{d\ln r} K_{\text{ext/sca}}^{\text{NS}}(\lambda, n, k, r_k), \quad (11)$$

$$\tau_{\text{sca}}(\lambda) P_{ii}(\Theta, \lambda) = \sum_k \frac{dV_f(r_k)}{d\ln r} K_{ii}^S(\Theta, \lambda, n, k, r_k) + \sum_k (1 - \delta) \frac{dV_c(r_k)}{d\ln r} K_{ii}^S(\Theta, \lambda, n, k, r_k) + \sum_k \delta \frac{dV_c(r_k)}{d\ln r} K_{ii}^{\text{NS}}(\Theta, \lambda, n, k, r_k), \quad (12)$$

where $\tau_{\text{ext/sca}}(\lambda)$ denotes the optical thickness of extinction and scattering, and $\tau_{\text{sca}}(\lambda) P_{ii}(\Theta, \lambda)$ denotes the directional scattering corresponding to the scattering matrix elements $P_{ii}(\Theta, \lambda)$. K^S and K^{NS} are the kernels of extinction and scattering properties for spherical and non-spherical particles, respectively. n and k are the real and imaginary parts of the refractive index, respectively. We use randomly oriented spheroids as non-spherical particles and use the kernels developed by Dubovik *et al.* (2006).

We compute normalized angular distributions and transmittances of the extinction, using the radiative transfer model RSTAR (Nakajima and Tanaka, 1986, 1988). The model atmosphere is divided by 18 boundary layers-altitudes of 0, 1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 15, 20, 30, 40, 50, 70, and 120 km. Atmospheric vertical profiles of temperature and pressure are obtained from the NCEP/NCAR Reanalysis 1 data. The absorption coefficients of H₂O, CO₂, O₃, N₂O, CO, CH₄, and O₂ are calculated by the correlated k -distribution method from the data table of Sekiguchi and Nakajima (2008).

The aerosol parameters for the best fit to all measurements (normalized angular distributions and transmittances at aerosol channels) and *a priori* information are obtained by minimizing the following cost function,

$$f(\mathbf{x}) = \frac{1}{2}(\mathbf{y}^{\text{meas}} - \mathbf{y}(\mathbf{x}))^T (\mathbf{W}^2)^{-1}(\mathbf{y}^{\text{meas}} - \mathbf{y}(\mathbf{x})) + \frac{1}{2}(\mathbf{y}_a(\mathbf{x}))^T (\mathbf{W}_a^2)^{-1}(\mathbf{y}_a(\mathbf{x})) \quad (13)$$

where vector \mathbf{y}^{meas} describes the measurements (normalized radiances R^{meas} and transmittances of total extinction T^{meas} ~~at aerosol channels~~) at the aerosol channels, vector \mathbf{x} describes the aforementioned aerosol parameters — $n(\lambda)$, $k(\lambda)$, C_i , and δ — to be estimated, vector $\mathbf{y}(\mathbf{x})$ comprises the values corresponding to \mathbf{y}^{meas} calculated from \mathbf{x} by the forward model (R^{ret} and T^{ret}), and matrix \mathbf{W}^2 is the covariance matrix of \mathbf{y} and is assumed to be diagonal. The diagonal elements of \mathbf{W} are ~~the~~ standard errors in the measurements. We set their values at 0.02 for T^{meas} , and 10% for R^{meas} .

To reduce the effects of observational error on retrieval and to conduct stable analyses, Dubovik and King (2000) considered restricting the spectral variability of the ~~refractive index~~ volume size distribution ~~and by~~ limiting the length of ~~the derivatives of~~ the refractive index derivative with respect to the wavelength ~~and the volume size distribution~~. They considered this *a priori* smoothness constraint as being of the same nature as a measurement and incorporated the smoothness constraint into their retrieval scheme. We also consider the smoothness constraints in this study. The second term of Eq. (13) consists of *a priori* information on the wavelength dependencies of the refractive index, aerosol optical thickness, and smoothness of the volume spectrum, which is described as

$$\mathbf{y}_a(\mathbf{x}) = (\mathbf{y}_a^{\text{Re}}, \mathbf{y}_a^{\text{Im}}, \mathbf{y}_a^{\text{Sca}}, \mathbf{y}_a^{\text{Abs}}, \mathbf{y}_a^{\text{Vol}})^T \quad (14)$$

where vectors \mathbf{y}_a^{Re} , \mathbf{y}_a^{Im} , $\mathbf{y}_a^{\text{Sca}}$, $\mathbf{y}_a^{\text{Abs}}$, and $\mathbf{y}_a^{\text{Vol}}$ are *a priori* information on the wavelength dependencies of the refractive index (real and imaginary parts), aerosol optical thickness (scattering and absorption parts), and smoothness of the volume spectrum, respectively. The matrix \mathbf{W}_a^2 in Eq. (13) is the covariance matrix for determining the strengths of the constraints.

We adapt the smoothness constraints of the second derivatives for the real and imaginary parts of the refractive index. The second derivatives are defined as

$$y_a^{\text{Re}(i)}(\mathbf{x}) = \left(\frac{\ln n(\lambda_i) - \ln n(\lambda_{i+1})}{\ln \lambda_i - \ln \lambda_{i+1}} - \frac{\ln n(\lambda_{i+1}) - \ln n(\lambda_{i+2})}{\ln \lambda_{i+1} - \ln \lambda_{i+2}} \right) \quad (15)$$

$$y_a^{\text{Im}(i)}(\mathbf{x}) = \left(\frac{\ln k(\lambda_i) - \ln k(\lambda_{i+1})}{\ln \lambda_i - \ln \lambda_{i+1}} - \frac{\ln k(\lambda_{i+1}) - \ln k(\lambda_{i+2})}{\ln \lambda_{i+1} - \ln \lambda_{i+2}} \right) \quad (16)$$

$$(i = 1, \dots, N_w - 2),$$

where $y_a^{\text{Re}(i)}$ and $y_a^{\text{Im}(i)}$ are the i -th elements of the vectors \mathbf{y}_a^{Re} and \mathbf{y}_a^{Im} , respectively. N_w is the number of wavelengths. The values entered into the weight matrix \mathbf{W}_a are 0.2 for the real part and 1.25 for the

imaginary part. These values are adopted from Dubovik and King (2000). Furthermore, we introduce the smoothness constraints to the spectral distributions of the scattering and absorption parts of the aerosol optical thickness by

$$y_a^{\text{Sca}(i)}(\mathbf{x}) = \left(\frac{\ln \tau_{\text{sca}}(\lambda_i) - \ln \tau_{\text{sca}}(\lambda_{i+1})}{\ln \lambda_i - \ln \lambda_{i+1}} - \frac{\ln \tau_{\text{sca}}(\lambda_{i+1}) - \ln \tau_{\text{sca}}(\lambda_{i+2})}{\ln \lambda_{i+1} - \ln \lambda_{i+2}} \right), \quad (17)$$

$$y_a^{\text{Abs}(i)}(\mathbf{x}) = \left(\frac{\ln \tau_{\text{abs}}(\lambda_i) - \ln \tau_{\text{abs}}(\lambda_{i+1})}{\ln \lambda_i - \ln \lambda_{i+1}} - \frac{\ln \tau_{\text{abs}}(\lambda_{i+1}) - \ln \tau_{\text{abs}}(\lambda_{i+2})}{\ln \lambda_{i+1} - \ln \lambda_{i+2}} \right), \quad (18)$$

$$(i = 1, \dots, N_w - 2),$$

where $y_a^{\text{Sca}(i)}$ and $y_a^{\text{Abs}(i)}$ are the i -th elements of the vectors $\mathbf{y}_a^{\text{Sca}}$ and $\mathbf{y}_a^{\text{Abs}}$, respectively. The value entered in the weight matrix \mathbf{W}_a is 2.5 for both the scattering and absorption parts of the aerosol optical thickness. To stabilize the estimation of the volume size distribution, we introduce the smoothness constraint for the adjacent volume size spectrum C_i , as:

$$y_a^{\text{Vol}(i)}(\mathbf{x}) = (\ln C_{i-1} - \ln C_i) - (\ln C_i - \ln C_{i+1}), \quad (19)$$

$$(i = 1, \dots, 20),$$

$$C_0 = 0.01 \times \min\{C_i | i = 1, \dots, 20\}, C_{21} = 0.01 \times \min\{C_i | r_i > r_b, i = 1, \dots, 20\}.$$

where $y_a^{\text{Vol}(i)}$ is the i -th element of the vector $\mathbf{y}_a^{\text{Vol}}$. The small values of C_0 and C_{21} at r_0 and r_{21} are given to prevent both ends of the size distribution (C_1 and C_{20}) from being abnormal values because ~~F the solar direct irradiances~~ and ~~V diffuse radiances~~ do not have sufficient information to estimate the size distribution of both small ($r < 0.1 \mu\text{m}$) and large particles ($r > 7 \mu\text{m}$; Dubovik *et al.*, 2000). Note that r_0 and r_{21} satisfy Eq. (7). The value entered in the weight matrix \mathbf{W}_a is 1.6 for the smoothness constraint of the size distribution.

We minimize $f(\mathbf{x})$ of Eq. (13) using the algorithm developed in Kudo *et al.* (2016), which is based on the Gauss-Newton method and the logarithmic transformations of \mathbf{x} and \mathbf{y} . Finally, the aerosol optical properties from aerosol channels are obtained from \mathbf{x} using Eqs. (11) and (12).

2.2.2 Step 2: Retrieval of PWV

We estimate PWV by the following procedure. The aerosol volume size distribution is obtained from step 1, and the refractive index at 940 nm is calculated from those at 870 and 1020 nm by linear interpolation in the log-log plane. Using the size distribution and the interpolated refractive index, we

can compute the aerosol optical properties and the normalized angular distribution at the water vapor channel using the forward model described in Section 2.2.1. We retrieve PWV by minimizing the following cost function:

$$f(\mathbf{x}) = \frac{1}{2}(\mathbf{y}^{\text{meas}} - \mathbf{y}(\mathbf{x}))^T (\mathbf{W}^2)^{-1}(\mathbf{y}^{\text{meas}} - \mathbf{y}(\mathbf{x})) \quad (20)$$

where the component of vector \mathbf{x} is PWV, vectors \mathbf{y}^{meas} and $\mathbf{y}(\mathbf{x})$ are the normalized angular distribution in the range ~~of~~ from 4° to 160° , matrix \mathbf{W}^2 is assumed to be diagonal, and the values of the diagonal matrix \mathbf{W} are assumed to be 10%. The cost function is minimized by the Gauss-Newton method. Note that this process does not require the calibration constants of the sky-radiometer, because we use the normalized angular distribution (Eq. [4]) to obtain PWV instead of using the direct solar irradiance (Eq. [1]).

2.2.3 Step 3: Retrieval of the calibration constant of the water vapor channel

F_0 at the water vapor channel can be obtained from the observed F and the band average transmittance $\bar{T}_{\text{H}_2\text{O}}$ converted from PWV in step 2 as follows:

$$F_0 = \frac{F d^2 e^{m \cdot (\tau_R + \tau_a)}}{\bar{T}_{\text{H}_2\text{O}}} \quad (21)$$

where τ_R and τ_a are Rayleigh scattering and aerosol optical thicknesses, respectively. The band average transmittance can be written as

$$\bar{T}_{\text{H}_2\text{O}} = \frac{\int_{\Delta\lambda} \Phi(\lambda) T_{\text{H}_2\text{O}}(\lambda) d\lambda}{\int_{\Delta\lambda} \Phi(\lambda) d\lambda} = \frac{\int_{\Delta\lambda} \Phi(\lambda) \exp(-m_{\text{H}_2\text{O}}(\theta) \int_0^z \alpha_{\text{H}_2\text{O}}(g_w(z), K(z), \lambda) dz) d\lambda}{\int_{\Delta\lambda} \Phi(\lambda) d\lambda} \quad (22)$$

$$w = \int_0^z g_w(z) dz \quad (23)$$

where $\Phi(\lambda)$ is the filter response function, $\Delta\lambda$ is the bandwidth of the filter response function, $T_{\text{H}_2\text{O}}$ is the transmittance of water vapor at wavelength λ , $m_{\text{H}_2\text{O}}(\theta)$ is the optical air mass, g_w is the mass mixing ratio, K is temperature, $\alpha_{\text{H}_2\text{O}}$ is the absorption coefficient at altitude z , and w is PWV. Eq. (22) is discretized by

$$\bar{T}_{\text{H}_2\text{O}} =$$

$$\frac{\sum_i^{N_s} \Phi_i \int_{\Delta\lambda_i} \exp(-m_{\text{H}_2\text{O}}(\theta) \int_0^z \alpha_{\text{H}_2\text{O}}(g_w(z), K(z), \lambda) dz) d\lambda}{\sum_i^{N_s} \Phi_i \Delta\lambda_i} \frac{1}{\Delta\lambda} \sum_i^{N_s} \Phi_i \int_{\Delta\lambda_i} \exp(-m_{\text{H}_2\text{O}}(\theta) \int_0^z \sigma_{\text{H}_2\text{O}}(g_w(z), K(z), \lambda) dz) d\lambda \quad (24)$$

where Φ_i is the stepwise filter response function, $\Delta\lambda_i$ is the sub-bandwidth of the filter response function, and N_s is the number of sub-bands. We calculate the absorption coefficients at each wavelength by the correlated k -distribution (Sekiguchi and Nakajima, 2008) using the vertical profiles of temperature, pressure, and specific humidity in the NCEP/NCAR Reanalysis 1 data.

We can calculate a value for F_0 from a data set of the normalized angular distribution. Therefore, for example, the time series of F_0 in a day is obtained from the daily measurements of the sky-radiometer. The mean value of the calibration constant at the water vapor channel is determined by the robust statistical and iterative method with Huber's M-estimation:

$$\ln \bar{F}_0 = \sum_i \beta_H(t_i) \cdot \ln F_0(t_i) \quad (25)$$

$$\beta_H(t_i) = \begin{cases} 1 & (|\ln \bar{F}_0 - \ln F_0(t_i)| \leq 0.03) \\ \frac{0.03}{|\ln \bar{F}_0 - \ln F_0(t_i)|} & (|\ln \bar{F}_0 - \ln F_0(t_i)| > 0.03) \end{cases} \quad (26)$$

where \bar{F}_0 is the mean calibration constant and is calculated at each iterative step, $F_0(t_i)$ is the calibration constant at a specific time t_i , and β_H is Huber's weight function.

2.2.4 Cloud screening using the smoothness criteria of the angular distributions (SCAD method)

The SKYMAP algorithm can only be applied to measurements under clear-sky conditions. We estimated clear-sky conditions from two indexes calculated from sky-radiometer measurements. Index 1 is a value for diffuse radiances the normalized radiances near the sun. If clouds pass over the sun, index 1 has large temporal variation. Index 2 is a value for the normalized angular distribution. If clouds are detected on the scanning plane of the sky-radiometer, the normalized angular distribution has large variation. Index 1 is defined as follows. First, the mean normalized radiance near the sun \bar{R}_{near} is calculated by

$$\bar{R}_{\text{near}}(t) = \frac{1}{N} \sum_{i=1}^N R(\theta_i, t), \theta \leq 10^\circ \quad (27)$$

where N is the number of measurements, and R is the normalized radiance at a time t , scattering angle Θ , and wavelength 500 nm. Next, the running mean of the time series of $\bar{R}_{\text{near}}(t)$ with a window of three consecutive data points is calculated as $\langle \bar{R}_{\text{near}}(t) \rangle$. Index 1 is defined as the deviation $\tilde{R}_{\text{near}}(t)$ of $\bar{R}_{\text{near}}(t)$ from $\langle \bar{R}_{\text{near}}(t) \rangle$.

$$\tilde{R}_{\text{near,dev}}(t) = |\bar{R}_{\text{near}}(t) - \langle \bar{R}_{\text{near}}(t) \rangle| / \langle \bar{R}_{\text{near}}(t) \rangle \quad (28)$$

Index 2 is the deviation \tilde{R}_{far} of normalized angular distributions far from the sun and is defined as

$$\tilde{R}_{\text{far,dev}}(t) = \sigma \left(\frac{R(\Theta, t) - \langle R_{\text{far}}(\Theta, t) \rangle}{\langle R_{\text{far}}(\Theta, t) \rangle} \right), \Theta > 10^\circ \quad (29)$$

where $\langle R_{\text{far}}(\Theta, t) \rangle$ is the running mean of $R(\Theta_i, t)$ with a window of three consecutive data points, and $\sigma(\mathbf{X})$ is the standard deviation of data set \mathbf{X} . Note that the data for calculating \tilde{R}_{far} varies depending on SZA, which limits available scattering angles. We judged clear-sky conditions, when two indexes 1 and 2 were both below their respective thresholds (0.1 and 0.2, respectively). We determined the thresholds by comparing the images of the whole-sky camera and the time series of the surface solar radiation observed by the pyranometer. Figure 9 is an example of the results for observations on January 6, 2014, in Tsukuba. Clear-sky conditions continued until 12:30, and then clouds passed over the sky until 15:00. Subsequently, there were clouds near the horizon, but the sky was almost clear. Our algorithm worked well, and cloudy scenes were eliminated. Although the whole-sky camera detected some clouds from 14:00 to 15:00, our algorithm judged the scenes as representative of clear-sky conditions. This may be because there were no clouds in the line of sight of the sky-radiometer. However, the cloudy conditions from 14:00 to 15:00 were misjudged, because the sky-radiometer observes only a part of the whole sky. The decline in the surface solar radiation around 9:00 was due to wiping of the glass dome of the pyranometer to keep the dome clean.

The method was applied to measurements from 2013 to 2014 at the Meteorological Research Institute, Japan Meteorological Agency (MRI, JMA), in Tsukuba. The results were validated using visual observation of the amount of clouds in the Aerological Observatory of the JMA. Figure 10a shows the histograms of index 1 for cases in which the sun was and was not covered by clouds. Index 1 had a low value when there were no clouds shading the sun but had a wide range of values when clouds were shading the sun. Fig. 10b shows the histograms of index 2 when cloud cover was and was not $\leq 20\%$. The peak shifted to the right when cloud cover was $\geq 20\%$, but the effect was not significant. Table 32 shows the validation results of this method. We defined “clear-sky best condition” as cloud cover $\leq 20\%$ and “cloud-affected poor condition” as cloud cover $\geq 20\%$. In less than 17% of cases a

“poor condition” cloudy sky was misjudged as a “best condition” clear sky. The sky-radiometer observes only a part of the whole sky, but our algorithm showed good results.

2.3 Estimation of PWV from direct solar irradiance (DSRAD algorithm)

The sky-radiometer observes the angular distribution of L_V every 10 min but observes the direct solar irradiance every 1 min. Once the calibration constant is determined by the SKYMAP algorithm, we can estimate PWV from the direct solar irradiance. The DSRAD algorithm computes the aerosol optical thickness, and PWV from the direct solar irradiances at the aerosol and water vapor channels. Table 43 shows the references of the DSRAD algorithm. This algorithm consists of two steps. First, aerosol optical thicknesses at aerosol channels are calculated using direct solar irradiances. The aerosol optical thickness at the water vapor channel is interpolated from the aerosol optical thicknesses at 870 and 1020 nm by linear interpolation in the log-log plane by line regression. Second, the band mean transmittance of the water vapor, $\bar{T}_{H_2O}^{meas}$, is calculated from the calibrated direct solar irradiance. PWV is retrieved using the formula,

$$\bar{T}_{H_2O}^{meas} = \frac{\sum_i^{N_s} \Phi_i \int_{\Delta\lambda_i} \exp(-m_{H_2O}(\theta) \int_0^z \alpha_{H_2O}(g_w(z), K(z), \lambda) dz) d\lambda}{\sum_i^{N_s} \Phi_i \Delta\lambda_i} \frac{\sum_i^{N_s} \Phi_i \int_{\Delta\lambda_i} \exp(-m_{H_2O}(\theta) \int_0^z \sigma_{H_2O}(g_w(z), K(z), \lambda) dz) d\lambda}{\sum_i^{N_s} \Phi_i \Delta\lambda_i} = 0. \quad (30)$$

where m_{H_2O} is the optical air mass calculated by Gueymard (2001). Eq. (30) is solved using the Newton–Raphson method.

To ensure the quality of the data and avoid cloud contamination, we adopt the method of Smirnov *et al.* (2000) with two main differences, similar to Estellés *et al.* (2012). First, an aerosol optical thickness at 500 nm > 2 is considered cloud-affected data. Second, the triplet of the aerosol optical thickness in Smirnov *et al.* (2000) is built from the pre/post 1 min data instead of 30 s.

3 Sensitivity tests using simulated data

We conducted sensitivity tests using simulated data to evaluate SKYMAP algorithm steps 1 and 2 (Figs. 7a and 7b). The simulation was conducted using the two aerosol types based on those used by Kudo *et al.* (2016): the continental average, and the continental average + transported dust in the upper atmosphere (Table 1). The continental average consisted of water-soluble particles, soot particles, and insoluble particles (Hess *et al.*, 1999) described in Section 2.1.2. Transported dust was defined as the mineral-transported component from Hess *et al.* (1999). The sensitivity test was conducted with sky

495 radiances in the almucantar plane for the wavelengths of 340, 380, 400, 500, 675, 870, 940, and 1020 nm; aerosol optical thicknesses of 0.02, 0.06, and 0.20 at 940 nm; PWV of 0.0, 0.5, 1.0, 1.5, 2.0, 2.5, 3.0, 3.5, 4.0, 4.5, and 5.0 cm; and SZA of 30°, 50°, and 70°.

Figure 101 illustrates the retrieval results from the simulated data for the continental average aerosol with aerosol optical thicknesses of 0.02, 0.06, and 0.20 at 940 nm. The retrievals of the volume size distribution, aerosol optical thickness, and PWV corresponded with their input values (“true” values in Fig. 11) when the input of PWV was <2 cm. This was seen regardless of the magnitude of the aerosol optical thickness. When the input of PWV was >2 cm, the volume size distribution, scattering and absorption optical thickness were retrieved well, but PWV was underestimated. When PWV was >2 cm, the normalized angular distribution was insensitive to PWV (Fig. 3). Figure 121 illustrates the retrieval results from the simulated data for the transported dust aerosol with aerosol optical thicknesses of 0.02, 0.06 and 0.20 at 940 nm. The scattering and absorption optical thicknesses were retrieved well. The volume size distribution of fine mode was slightly overestimated. The rRetrieval errors of PWV increased with increasing aerosol optical thickness because the near-infrared wavelength was strongly affected by the retrieval of coarse mode particles.

510 We also conducted sensitivity tests using the simulated data with bias errors to investigate uncertainty in the SKYMAP-derived PWV. The bias errors were $\pm 5\%$ and $\pm 10\%$ for R . The value of 5% was given by following reasons. The SVA bias errors of the diffuse radiances for the sky-radiometer observations were estimated to be less than 5% (Uchiyama *et al.*, 2018b). According to Dubovik *et al.* (2000), the uncertainty of the diffuse radiances for the AERONET measurements is $\pm 5\%$. Figures 13 and 14 show the results from the simulated data for the continental average and transported dust aerosols with aerosol optical thicknesses of 0.02, 0.06 and 0.20 at 940 nm. PWV was overestimated when -5% bias was applied to R . This corresponds to the relationship between R and PWV, where R decreases with increasing PWV (Section 2.1.2). The bias errors strongly affected the retrieval of PWV at high PWV (> 2 cm), because the sensitivity of high PWV is lower than that of low PWV. The retrieval error of PWV increased with increasing bias errors. The retrieval error of PWV due to $\pm 5\%$ and $\pm 10\%$ errors for R was within 10% for PWV < 2 cm and up to 200% for PWV > 2 cm.

525 When the input of PWV was < 2 cm, the SKYMAP algorithm retrieved PWV very well, within an error of 0.5–em10%, regardless of the aerosol optical thickness or the aerosol type. This was also observed when the bias errors were added for R . The scattering and absorption parts of the aerosol optical thickness were also estimated very well within ± 0.01 in all conditions. Present sensitivity tests suggest the design of a sky-radiometer calibration program as follows: to determine the calibration constant of the water vapor channel in dry days/seasons with PWV < 2 cm, and to obtain PWV from direct solar irradiance data throughout the year, as illustrated in Fig. 1.

4 Application to observational data

530 We applied our methods to SKYNET sky-radiometer data in Tsukuba and Chiba. The results were compared to PWV observed by well-established instruments and methods other than the sky-radiometer. ~~Aerosol~~ The aerosol channels of the sky-radiometer were calibrated by the IL method with SKYRAD.pack version 4.2 (Nakajima *et al.*, 1996; Campanelli *et al.*, 2004, 2007), and the solid view angles of all channels were calibrated by the on-site methods (Nakajima *et al.*, 1996; Boi *et al.*, 1999; 535 Uchiyama *et al.*, 2018b).

4.1 Observation at Tsukuba

In Tsukuba, the sky-radiometer model POM-02 (S/N PS1202091) is installed at the MRI (36.05°N, 140.12°E). We used data from 2013 and to 2014. The water vapor channel of PS1202091 was calibrated each winter by side-by-side comparison with the reference sky-radiometer, which was calibrated by the 540 standard Langley method at the NOAA Mauna Loa Observatory (Uchiyama *et al.*, 2014). PWV was also observed using a GNSS/GPS receiver (Shoji, 2013) at Ami station (No. 0584; 36.03°N, 140.20°E), approximately 7.5 km east-southeast of the MRI.

The calibration constant of the water vapor channel was determined for each month (Figs. 152a and 163a). To obtain the correct value, we used the retrieval results with $PWV_{SKYMAP} < 2$ cm and 545 sufficiently small cost functions (Eqs. [13]) and [20]). The annual mean values-calibration constants for 2013 and 2014 were 1.886×10^{-4} A and 2.212×10^{-4} A, respectively. The annual mean calibration constants changed drastically from 2013 to 2014 (+ 17.2%). This is Bbecause the lens at the visible and near-infrared wavelengths was replaced in December 2013, ~~the calibration constants at these wavelengths changed drastically (annual mean value: + 17.2 % from 2013 to 2014).~~ These results in 550 2013 and 2014 were ~~less,~~ -10.1% and -3.2% lower, respectively, than those determined by the side-by-side comparison with the reference sky-radiometer. The difference in the value of the calibration constant between the SKYMAP algorithm and the side-by-side comparison with the reference sky-radiometer was attributable mainly to the calibration period. The calibration constant of the sky-radiometer has seasonal variation due to the temperature dependency of the sensor output (Uchiyama *et al.*, 2018a). Calibration by side-by-side comparison with the reference sky-radiometer was performed only in the winter. However, the calibration constant of the SKYMAP algorithm was the annual mean.

~~Although the monthly mean calibration constant of the water vapor channel was underestimated every year in the wet season (May to October), it was a good estimate in the dry season (November to April). The number of retrieved results was small in summer because of cloudiness. In summer in Japan, 560 clouds develop every day because it is warm with high relative humidity. Thus, because of higher aerosol optical thickness and as a result of the cloud-affected data, PWV in summer also contained large bias. We rejected the monthly calibration constant, which was calculated from fewer 50 data sets.~~

Figures 152b and 163b show the DSRAD-retrieved PWV, which is denoted by $PWV_{DSRAD+SKYMAP}$, using the monthly calibration constant. $PWV_{DSRAD+SKYMAP}$ of the sky-radiometer agreed well with that of the GNSS/GPS receiver. Note that we did not retrieve PWV using the monthly mean calibration constants for June and July 2014 because their values were obviously small, and because little data were successfully retrieved due to the wet and cloudy conditions in the summer. In addition, it is possible that the measurements were contaminated by clouds.

Although monthly mean calibration constants values are best, in theory, they could not be obtained during the wet season or during periods of high aerosol optical thickness ~~in~~ due to the transported dust. Thus, we ~~calculated-used~~ the annual mean calibration constant ~~value~~ from all data in a year to estimate PWV. Figures 152c and 163c ~~describe-illustrate~~ PWV using the annual mean ~~value~~ calibration constants ~~for the year~~. The retrieved PWV agreed well with PWV from the GNSS/GPS receiver (correlation coefficient $\gamma = 0.987$ and 0.987 , and slope = 0.919 and 0.934 for 2013 and 2014, respectively; Table 5). We estimated PWV, which is denoted by $PWV_{DSRAD+LM}$, from the DSRAD algorithm using the calibration constant obtained by the side-by-side comparison with the reference sky-radiometer. The comparison of $PWV_{DSRAD+LM}$ and the GNSS/GPS-derived PWV in Figs. 12d and 13d shows ~~the~~ good agreement, and the results are similar to those in Figs. 152c and 163c. Then we compared $PWV_{DSRAD+LM}$ and $PWV_{DSRAD+SKYMAP}$ in Figs. 152e and 163e. The difference between $PWV_{DSRAD+LM}$ and $PWV_{DSRAD+SKYMAP}$ was small: 17% in 2013, and 8% in 2014. Our self-calibration method showed comparable results to those based on the standard Langley method (*Uchiyama et al., 2014*). Table 5 summarizes the results of comparisons of DSRAD-derived PWV and GNSS/GPS-derived PWV. The magnitude of the bias error and root mean square error were small, less than 0.11 cm and less than 0.226 cm, during 2013 to 2014. Table 6 shows the errors of the retrieved PWV with the annual mean calibration constants for the rank of PWV. The bias error was larger for high PWV than it was for low PWV. The magnitude of the bias errors of PWV was less than 0.163 cm for $PWV < 3$ cm and less than 0.339 cm for $PWV > 3$ cm.

4.2 Observation at Chiba

We used ~~2017-the~~ data from the sky-radiometer model POM-02 (S/N PS2501417) at Chiba University (35.63°N, 140.10°E) in 2017. ~~The~~ PWV was also obtained by a Radiometrix MP-1500 microwave radiometer (MWR) and AERONET sun-sky radiometer (Cimel, France) at the same location. The MWR measured in the 22-30 GHz region at 1-min temporal resolution and retrieved PWV_{MWR} using default software. PWV_{Cimel} of the AERONET sun-sky radiometer was retrieved by the direct solar irradiance at ~~93640~~ nm with adjustment parameters (direct sun algorithm version 3; Holben *et al.*, 1998; Giles *et al.*, 2018) and adopted the cloud screening method (AERONET Level 2.0). The AERONET product comprises three types of data: Level 1.0 data are not screened for cloud-affected or low-quality

-data, Level 1.5 data are screened but not completely calibrated, and Level 2.0 data are finalized data that have been calibrated and screened. We used PWV for the Level 2.0 data.

Figure 174 shows comparisons of $PWV_{DSRAD+SKYMAP}$ using ~~the~~ monthly and annual means ~~of~~ ~~the~~ calibration constants, PWV_{MWR} , and PWV_{Cimel} . $PWV_{DSRAD+SKYMAP}$ using monthly means calibration constants agreed well (correlation coefficient $\gamma = 0.961$ and slope = 0.964) with those of the MWR (Fig. 174b). $PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant agreed with PWV_{MWR} (Fig. 17c). ~~within ± 0.05 cm~~ The error of $PWV_{DSRAD+SKYMAP}$ was $-0.041 < \text{bias} < 0.024$ cm and $RMSE < 0.212$ cm for low PWV (< 3 cm), ~~but and was $\text{bias} < -0.356$ cm and $RMSE > 0.465$ cm~~ smaller than PWV_{MWR} for high PWV (Table 6). ~~(Fig. 14e).~~ Figure 174d ~~also~~ shows that $PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant also agreed with PWV_{Cimel} for low PWV (< 3 cm) but was smaller than PWV_{Cimel} ~~when for high PWV was (> 3 cm).~~ PWV_{MWR} was larger than PWV_{Cimel} (Fig. 174e). $PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant was 12% and 9.1% smaller than PWV_{MWR} and PWV_{Cimel} , ~~but the difference was 10 % compared to the MWR and 16% compared to the AERONET sun-sky radiometer, respectively (Table 5).~~ These results suggest an underestimation of $PWV_{DSRAD+SKYMAP}$, ~~as the uncertainty of PWV_{Cimel} compared to the GNSS/GPS receiver is expected to be less than 10% (Giles et al., 2018).~~ The underestimation of $PWV_{DSRAD+SKYMAP}$ was due to two factors. The first is the retrieval of PWV by the annual mean calibration constant for the water vapor channel. The calibration constant not only is subject to aging but also undergoes seasonal variation due to temperature dependency (Uchiyama et al., 2018a). Thus, it is possible to underestimate the calibration constant in the wet season. Second, uncertainty regarding the aerosol optical thickness affected PWV retrieval. Figure 18 depicts the differences in PWV and aerosol optical thicknesses at 675, 870, and 1020 nm between the DSRAD algorithm and the AERONET retrieval. In the periods from January to May and from October to November, the differences in PWV and aerosol optical thicknesses were less than 0.1 cm and 0.015, ~~respectively.~~ However, the difference in PWV was greater than 0.1 cm from July to September. This corresponds to the difference in aerosol optical thicknesses at 675, 870, and 1020 nm from July to September, which indicates that the transmittance of water vapor was overestimated by the overestimation of aerosol optical thickness. This led to the underestimation of $PWV_{DSRAD+SKYMAP}$ using the annual mean calibration constant when PWV was > 3 cm. In our error estimation, the error of $+ 0.03$ for the aerosol optical thickness at 940 nm resulted in the error of $- 0.214$ cm for PWV (Appendix B).

5 Summary

We developed a new on-site self-calibration method, SKYMAP, to retrieve PWV from sky-radiometer data at the water vapor channel. This method first retrieves PWV from the normalized angular distribution ~~of the normalized radiance~~ without the calibration constant. Then the calibration constant is retrieved from the obtained PWV. Once the calibration constant is determined, PWV can be estimated

from the direct solar irradiance. Our DSRAD algorithm retrieves PWV from the direct solar irradiance. This method does not require ~~any of the~~ adjustment parameters used in the empirical methods of previous studies (e.g., Holben *et al.*, 1998; Uchiyama *et al.*, 2014; Campanelli *et al.*, 2014, 2018). Instead, the filter response function and the vertical profiles of water vapor, temperature, and pressure are required as input parameters. Thus, our physics-based algorithm has the potential to be applied to sky-radiometers all over the world. This is the greatest advantage of the present study.

Sensitivity tests using simulated data from sky-radiometer measurements showed that the SKYMAP algorithm retrieved PWV within an error of ~~0.5–cm~~ 10% for cases when PWV was <2 cm. ~~Much more~~ Larger retrieval errors occurred; ~~when in the cases when~~ PWV was >2 cm; because PWV became less sensitive to the normalized angular distribution ~~of the normalized radiance~~. Therefore, the SKYMAP algorithm can be applied only to dry conditions.

We applied SKYMAP and DSRAD algorithms to the sky-radiometer measurements at two SKYNET sites (Tsukuba and Chiba, Japan). At Tsukuba, the calibration constant estimated by the SKYMAP algorithm was compared to that obtained by side-by-side comparison with ~~a the reference~~ sky-radiometer calibrated by the standard Langley method. ~~Their differences were~~ The calibration constant calculated by the SKYMAP algorithm was –10.1% lower in 2013 and –3.2% lower in 2014 compared with the calibration constant estimated by side-by-side comparison. Our retrieved PWV data were compared to those obtained by a GNSS/GPS receiver, a microwave radiometer, and an AERONET sun-sky radiometer. The correlation coefficients and slopes were as good as >0.96 and 1.00 ± 0.12 , respectively. The magnitude of the bias error and the root mean square error were < 0.163 cm and < 0.251 cm, respectively, for low PWV (< 3 cm). However, our retrieved PWV was underestimated in the wet conditions, and the magnitude of the bias error and the root mean square error were less than 0.594 cm and less than 0.722 cm for high PWV. This was due to seasonal variation in the calibration constant and the overestimation of aerosol optical thickness at 940 nm interpolated from those at 870 and 1020 nm.

These results show that our new on-site self-calibration method is practical. In future work, we plan to compare our method with others in the SKYNET framework (Uchiyama *et al.* 2014; Campanelli *et al.*, 2014).

7

6 Data availability

The SKYMAP and DSRAD algorithms are available on request from the first author. The sky-radiometer data are available from the SKYNET website (<http://www.skynet-isdc.org/>), but the sky-radiometer data in Tsukuba, Japan, are available on request from the first author. The MWR data at Chiba University are available from CEReS, Chiba University (<http://atmos3.cr.chiba-u.jp/skynet/>). The

665 AERONET sun-sky radiometer data are available from the AERONET website
(<https://aeronet.gsfc.nasa.gov/>).

Author contributions

This study was designed by MM, RK, KA, TM, KM, and TN. Sky-radiometer measurements at Tsukuba were conducted by RK. Sky-radiometer and MWR measurements at Chiba were conducted by
670 HO and HI. Analyses of both sky-radiometers were performed by MM. The calibration constant of the sky-radiometer by the Langley method was provided by AU. Analyses of the GPS receiver were conducted by YS. Visual observations at Tsukuba were conducted by OI and MT. The manuscript was written by MM and RK, and all authors contributed to editing and revision.

Competing interests

675 The authors declare that they have no conflict of interest.

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685 accessed September 2018).

Appendix A: Width of the volume size distribution

Because $\frac{dV(r)}{d\ln r}$ is expressed by the superposition of 20-modal lognormal size distributions (Eq. [6]), the
width of $\frac{dV(r)}{d\ln r}$ is larger than that of each lognormal size distribution. The width of the lognormal size
distribution should be small to deal with the complicated and step variations in $\frac{dV(r)}{d\ln r}$. However, $\frac{dV(r)}{d\ln r}$
690 cannot represent a natural curve if η is large and s is small (Fig. A1). Hence, we have to find the
maximum value of η for making $\frac{dV(r)}{d\ln r}$ a natural curve. When C_i is constant, such value of η minimizes

the roughness of $\frac{dV(r)}{d\ln r}$, and $\frac{dV(r)}{d\ln r}$ approaches to a flat shape. For a simple formulation, we consider the function $A(x)$ which consists of the multimodal normal distribution function B_i with a constant height. $A(x)$ and B_i are expressed as

$$A(x) = \sum_{i=-\infty}^{\infty} B_i(x) = \sum_{i=-\infty}^{\infty} \exp \left[-\frac{\eta^2}{2} \left(\frac{x-i\xi}{\xi} \right)^2 \right], \quad (\text{A1})$$

where $i\xi$ and $\frac{\xi}{\eta}$ are the mean and standard deviation, respectively. Its differential is written as

$$\frac{dA}{dx} = \sum_{i=-\infty}^{\infty} \frac{dB_i}{dx} = \sum_{i=-\infty}^{\infty} -\eta^2 \left(\frac{x-i\xi}{\xi} \right) \exp \left[-\frac{\eta^2}{2} \left(\frac{x-i\xi}{\xi} \right)^2 \right]. \quad (\text{A2})$$

When the shape of $A(x)$ approaches to be flat, the difference between local maximum and minimum values of $A(x)$ is approximately 0. Because $\frac{dB_i}{dx}$ equals 0 at $x = j\xi$ ($j \in \mathbb{Z}$), $A(x)$ has the local maximum and minimum at $x = j\xi$ and $\left(j + \frac{1}{2}\right)\xi$ in $j \leq \frac{x}{\xi} < j + 1$. The difference Δ between the local maximum and minimum values is obtained as

$$\Delta = 1 - \frac{A\left(\frac{2j+1}{2}\xi\right)}{A(j\xi)}. \quad (\text{A3})$$

Figure A2 shows the relation between η and Δ . The value of Δ increases drastically at around $\eta = 1.5$. in addition, the shape of $\frac{dV(r)}{d\ln r}$ is unnatural when $\eta = 2.0$ (Fig. A1). Therefore, the value of η should be selected from the values around $\eta = 1.5$. In this study, we fixed η at 1.65. This value represents the natural curve of $\frac{dV(r)}{d\ln r}$ and satisfies that the value of Δ is small enough, $\Delta = 3.0 \times 10^{-3}$.

Appendix B: Error propagation from aerosol optical thickness to PWV

We evaluated the influence of the uncertainty of aerosol optical thickness on PWV using the empirical equation of Bruegge *et al.* (1992). PWV is described using the adjustment parameters as follows

$$w = \frac{1}{m_0} \left(-\frac{\ln \bar{T}_{\text{H}_2\text{O}}}{a} \right)^{\frac{1}{b}} [\text{cm}]. \quad (\text{B1})$$

The uncertainty of PWV ϵ_{PWV} is given from the partial differentiation of Eq. (B1) with respect to $\ln \bar{T}_{\text{H}_2\text{O}}$ as follows

$$\epsilon_{\text{PWV}} = \frac{\partial w}{\partial \ln \bar{T}_{\text{H}_2\text{O}}} \epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}} = \frac{w}{b \ln \bar{T}_{\text{H}_2\text{O}}} \epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}} \quad (\text{B2})$$

where $\epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}}$ is the uncertainty of $\bar{T}_{\text{H}_2\text{O}}$. Using Eq. (B1) with the adjusting parameters of the sky-radiometer, with $a = 0.620$ and $b = 0.625$ as the coefficient values for the trapezoidal spectral response function (Uchiyama *et al.*, 2018a), we write the uncertainty of PWV as

$$\epsilon_{\text{PWV}} = -\frac{w}{ab} (m_0 w)^{-b} \epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}} = -\frac{w}{0.388} (m_0 w)^{-0.625} \epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}} \quad (\text{B3})$$

If the uncertainty of the calibration constant at the water vapor channel is ignored, the uncertainty of $\bar{T}_{\text{H}_2\text{O}}$ is given from Eq. (21) as follows

$$\epsilon_{\ln \bar{T}_{\text{H}_2\text{O}}} = m_0 \epsilon_{\text{AOT}} \quad (\text{B4})$$

where ϵ_{AOT} is the uncertainty of the aerosol optical thickness at 940 nm. The uncertainty of PWV is written by Eqs. (B3) and (B4) as

$$\epsilon_{\text{PWV}} = -\frac{1}{0.388} (m_0 w)^{0.375} \epsilon_{\text{AOT}} = -0.214 [\text{cm}] \quad (\text{B5})$$

where $m_0 = 3.0$, $w = 5.0 \text{ cm}$, and $\epsilon_{\text{AOT}} = 0.03$.

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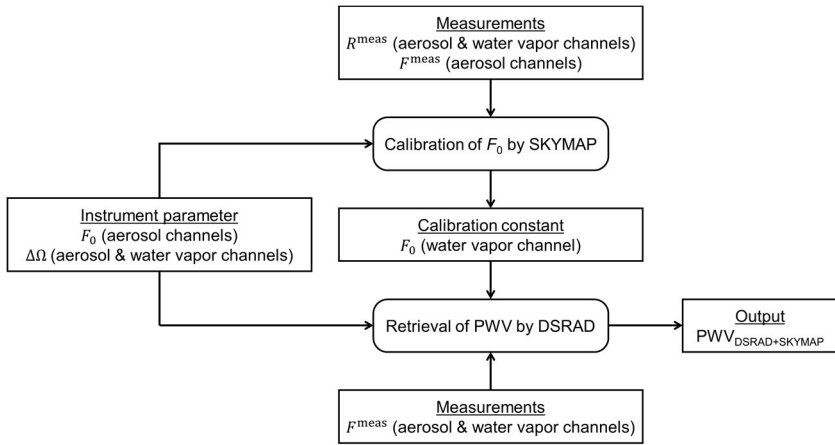


Figure 1: Diagram of the on-site self-calibration method (SKYMAP) and retrieval of PWV from direct solar irradiances (DSRAD). square boxes show the operation of the calculation and input/output parameters and rounded boxes show the operation of the algorithm.

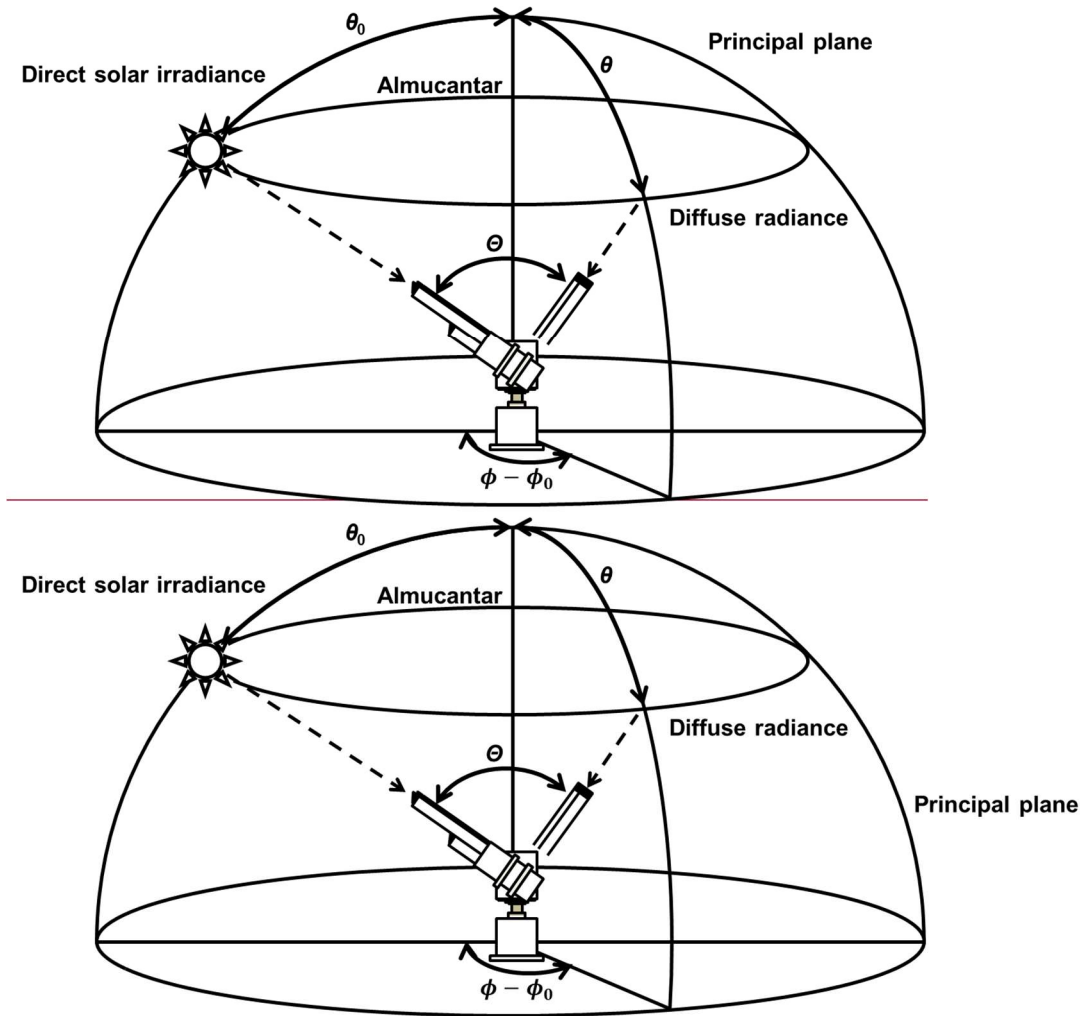


Figure 2: Observation planes (almucantar and principal planes) of the sky-radiometer.

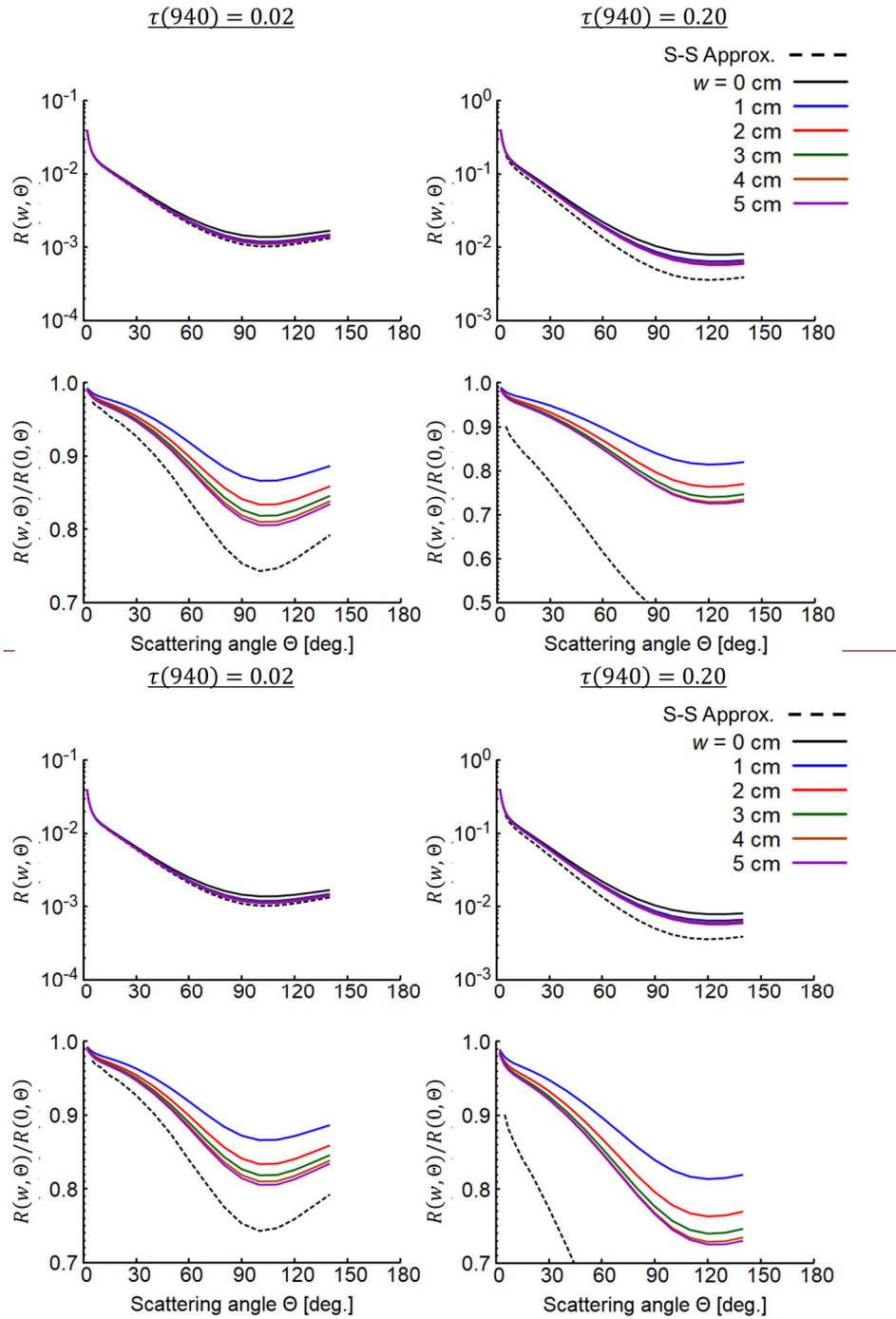


Figure 3: Normalized angular distributions simulated for continental average aerosol (Table 21) in the almucantar plane with aerosol optical thicknesses of 0.02 and 0.20 at 940 nm. The simulations were conducted for SZA = 70° and PWV (w) = 0, 1, 2, 3, 4, and 5 cm. The top row is the normalized radiance $R(w, \theta)$, and the bottom row is the ratio of $R(w, \theta)$ to $R(0, \theta)$. S-S Approx. is single scattering approximation.

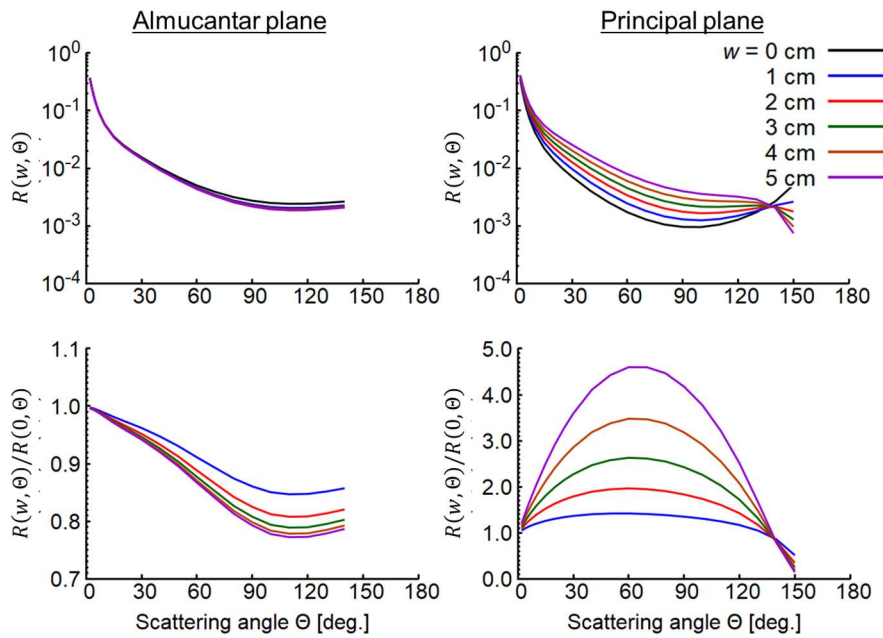


Figure 4: Normalized angular distributions simulated for transported dust aerosol (Table 21) in the almucentar and principal planes withat an aerosol optical thickness of 0.06 at 940 nm. The sSimulations were conducted for SZA = 70° and PWV (w) = 0, 1, 2, 3, 4, and 5 cm. The top rowline is the normalized radiance $R(w, \Theta)$, and the bottom rowline is the ratio of $R(w, \Theta)$ to $R(0, \Theta)$.

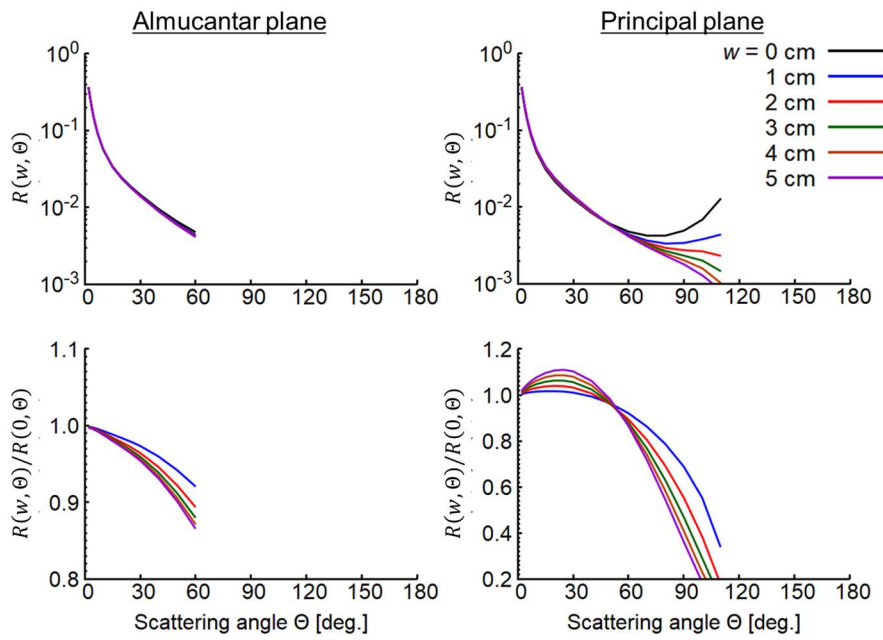


Figure 5: Similar to Fig. 4 but for SZA = 30°.

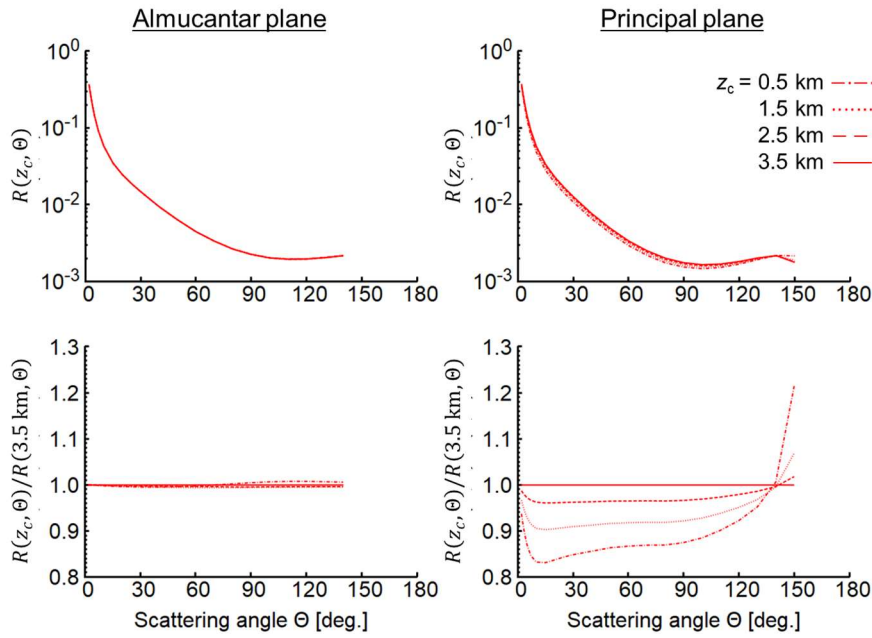
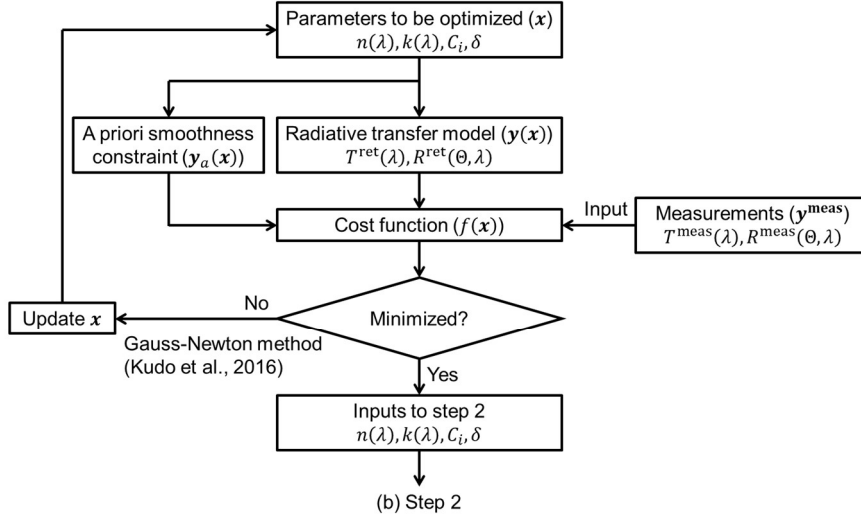
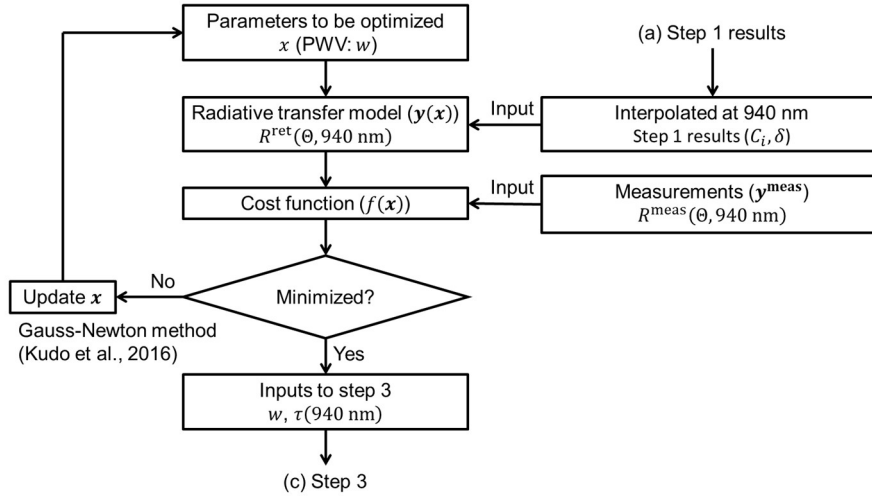


Figure 6: Normalized angular distributions simulated for transported dust aerosol (Table 21) in the almucantar and principal planes withat an aerosol optical thickness of 0.06 at 940 nm. The sSimulations were conducted for SZA = 70° and PWV = 2 cm. The height of the dust layer (z_c) is changed to 0.5, 1.5, 2.5, and 3.5 km. The top rowline is the normalized radiance $R(z_c, \Theta)$, and the bottom rowline is the ratio of $R(z_c, \Theta)$ to $R(3.5 \text{ km}, \Theta)$.

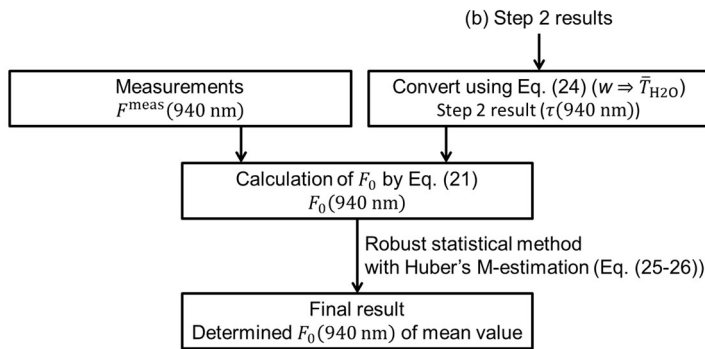
(a) Step 1: Retrieval of aerosol optical and microphysical properties



(b) Step 2: Retrieval of PWV



(c) Step 3: Retrieval of the calibration constant of the water vapor channel



900 **Figure 7: Schematic diagrams of SKYMAP procedures. (a) Step 1. (b) Step 2. (c) Step 3. Square boxes show the calculation and input/output parameters**

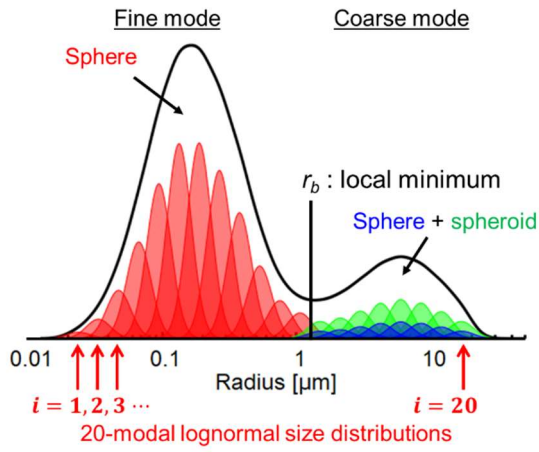


Figure 8: Assuming volume size distributions in the SKYMAP algorithm. Fine and coarse mode particles are separated at radius r_b . Spheroid particles are assumed only in coarse mode. The black line is the volume size distribution, which is computed by the integration of 20-modal lognormal distribution functions (red, blue, and green lines).

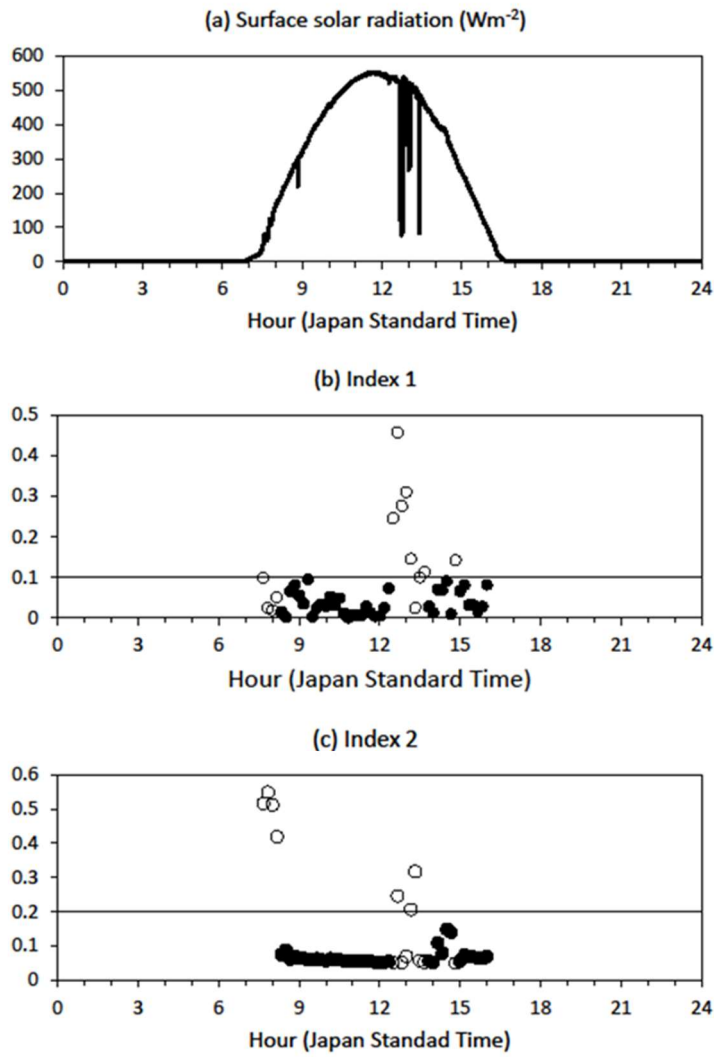


Figure 9: An example result of the SCAD method on January 6, 2014, in Tsukuba. (a) Surface solar radiation observed by the pyranometer. (b) Index 1. (c) Index 2. The closed circles indicate clear-sky conditions and the open circles indicate cloudy conditions in (b) and (c). The lines at 0.1 in (b) and 0.2 in (c) are thresholds for indexes 1 and 2, respectively.

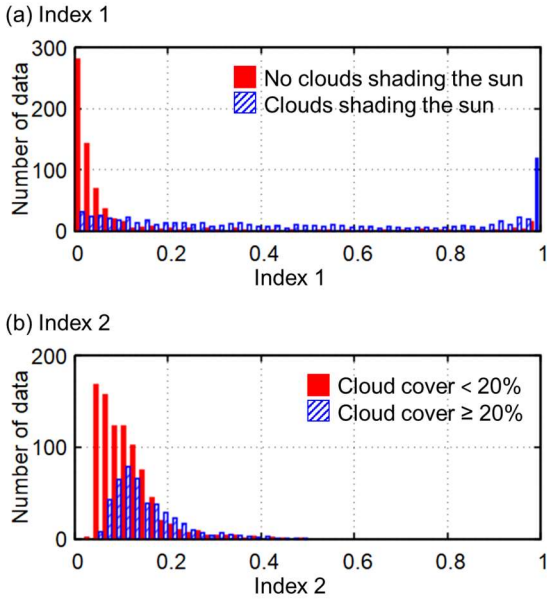


Figure 10: Histograms of indexes 1 and 2 of sky-radiometer observations at Tsukuba. (a) Index 1 when the sun is covered by clouds (blue boxes) and not covered by clouds (red boxes). (b) Index 2 when cloud cover is less than to 20% (red boxes) and greater than or equal to 20% (blue boxes).

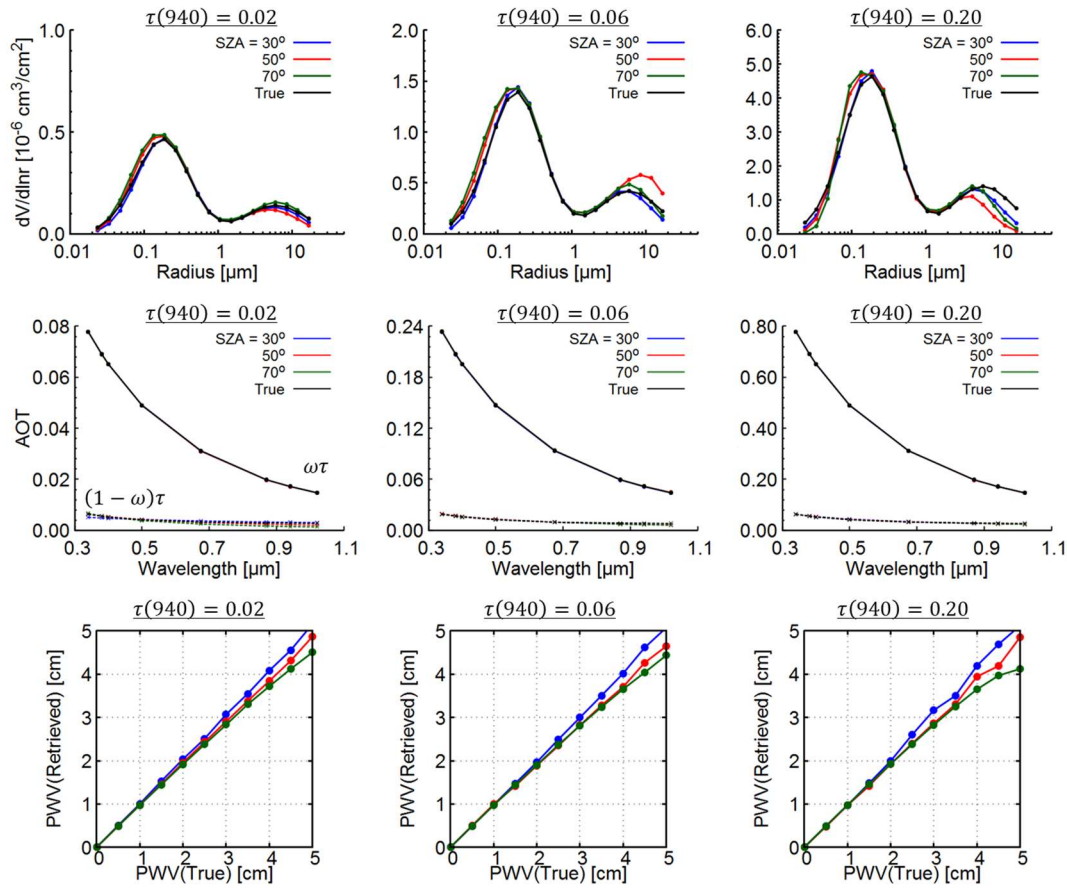


Figure 11: Retrieval results from simulated data for continental average aerosol. The top row is the volume size distribution, the middle row is the scattering and absorption parts of aerosol optical thickness, and the bottom row is a comparison of the “true” and retrieval values of PWV. Blue, red, and green lines are the retrieval results at SZA = 30°, 50°, and 70°.

respectively. The black line is the “true” value. Note that the blue, red, green, and black lines in the middle row overlap.

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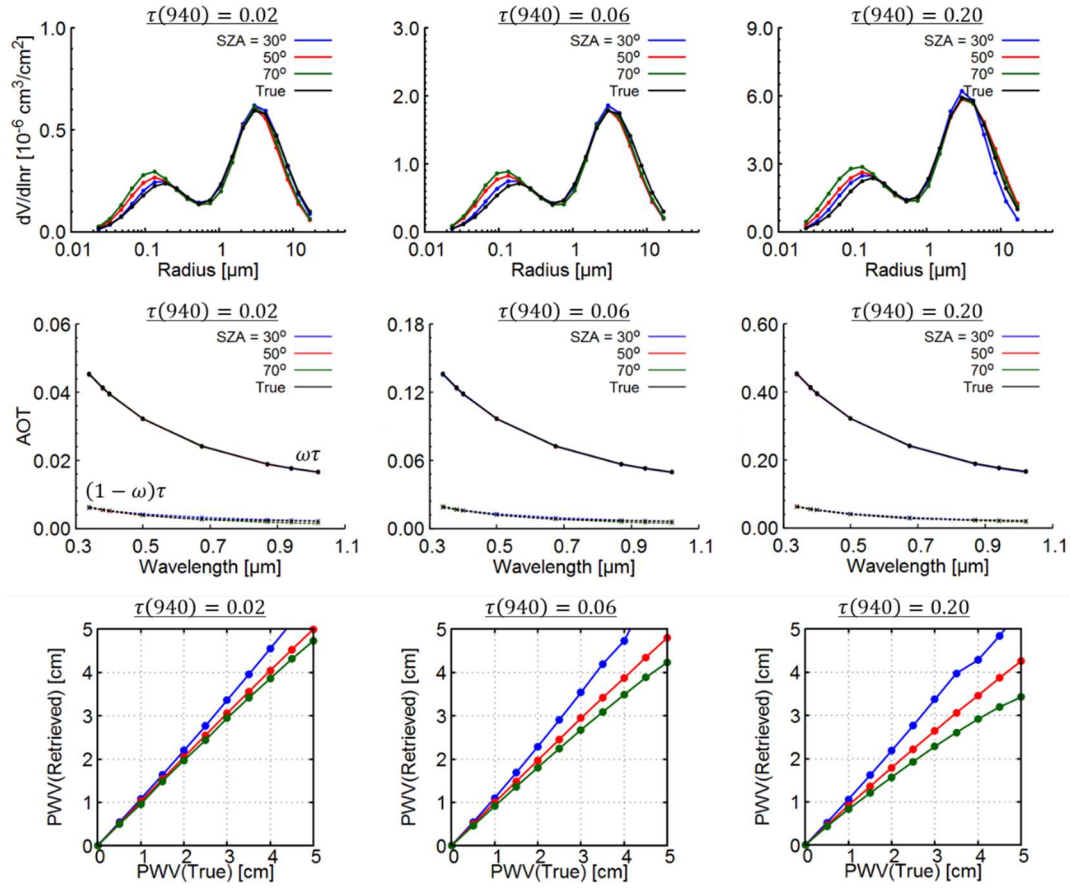


Figure 1211: Similar to Fig. 110 but for transported dust aerosol. Note that the blue, red, green, and black lines in the middle row overlap.

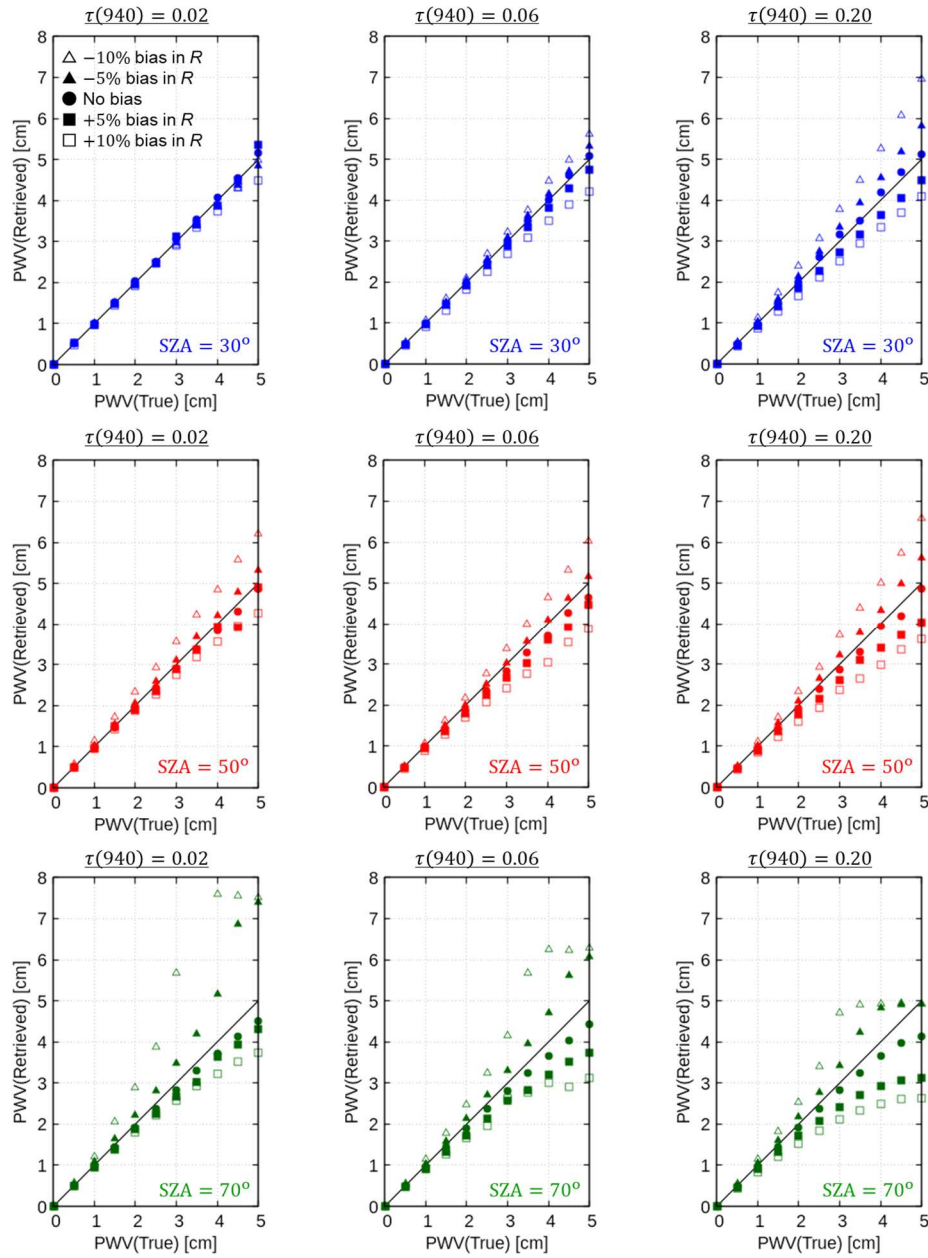


Figure 13: Comparison of the “true” and retrieval values of PWV from simulated data for continental average aerosol with bias errors. The top, middle, and bottom rows are the retrieval results at SZA = 30°, 50°, and 70°, respectively. Closed circles are the results with no bias errors. Closed squares and closed triangles are the results with bias errors of plus and minus 5% in R , respectively. Open squares and open triangles are the results with bias errors of plus and minus 10% in R , respectively.

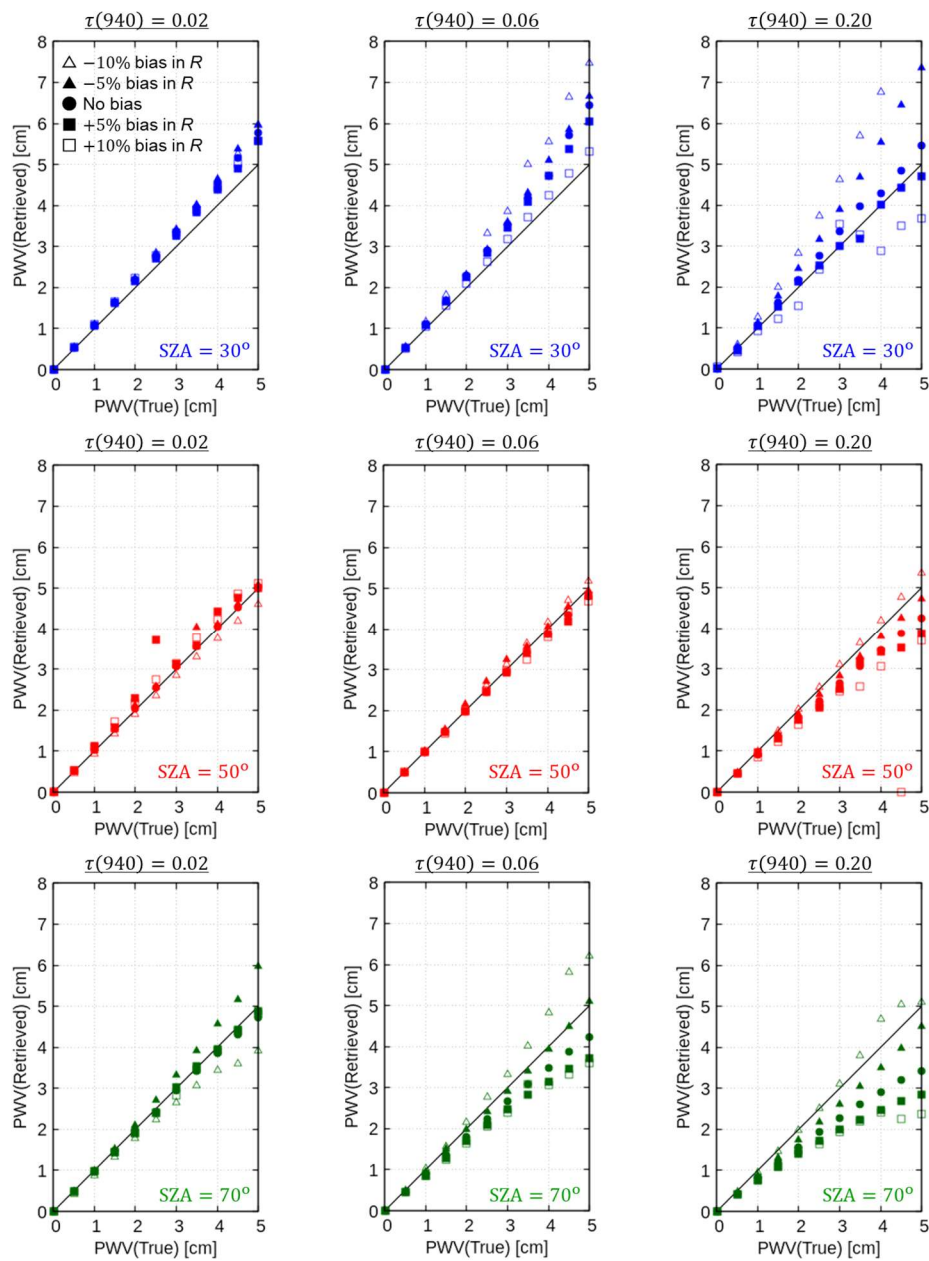


Figure 14: Similar to Fig. 13 but for transported dust aerosol.

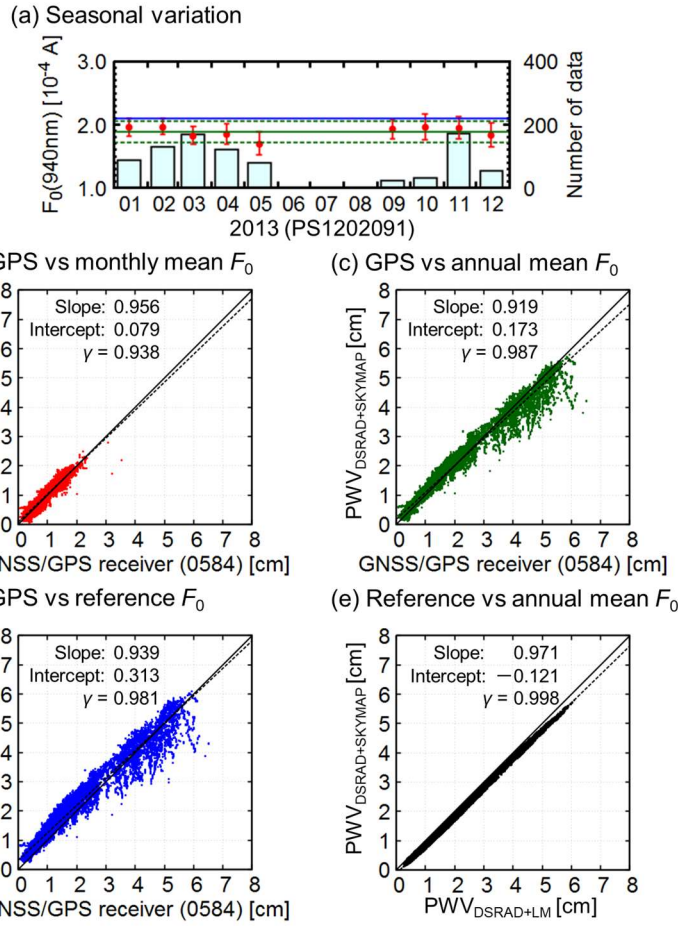


Figure 1512: Application of our methods to observational data from Tsukuba in 2013. (a) Seasonal variation in the calibration constant of the water vapor channel (red circles and error bars are monthly means and standard deviations, respectively; green solid and dotted lines are annual means and standard deviations, respectively; the blue line is the value obtained by a side-by-side comparison with the reference sky-radiometer; boxes indicate the number of data points). (b-d) Comparisons of PWV between the GNSS/GPS receiver and the sky-radiometer with (b) the monthly mean F_0 , (c) the annual mean F_0 , and (d) the reference F_0 . (e) Comparison of PWV from the sky-radiometer with the reference and annual mean F_0 .

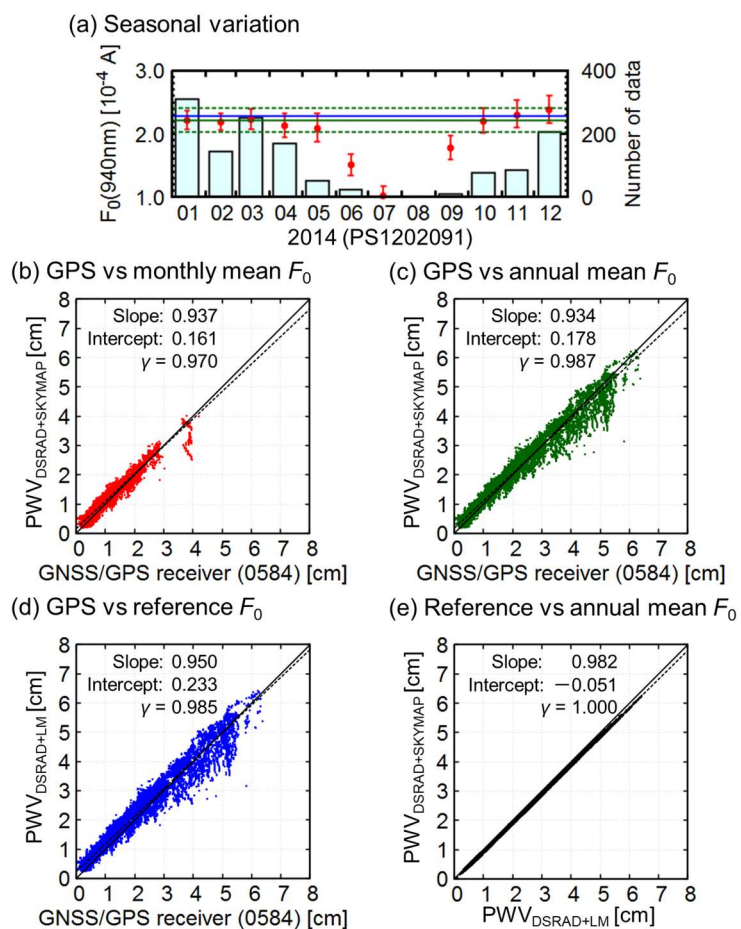


Figure 1613: Similar to Fig. 152 but in 2014.

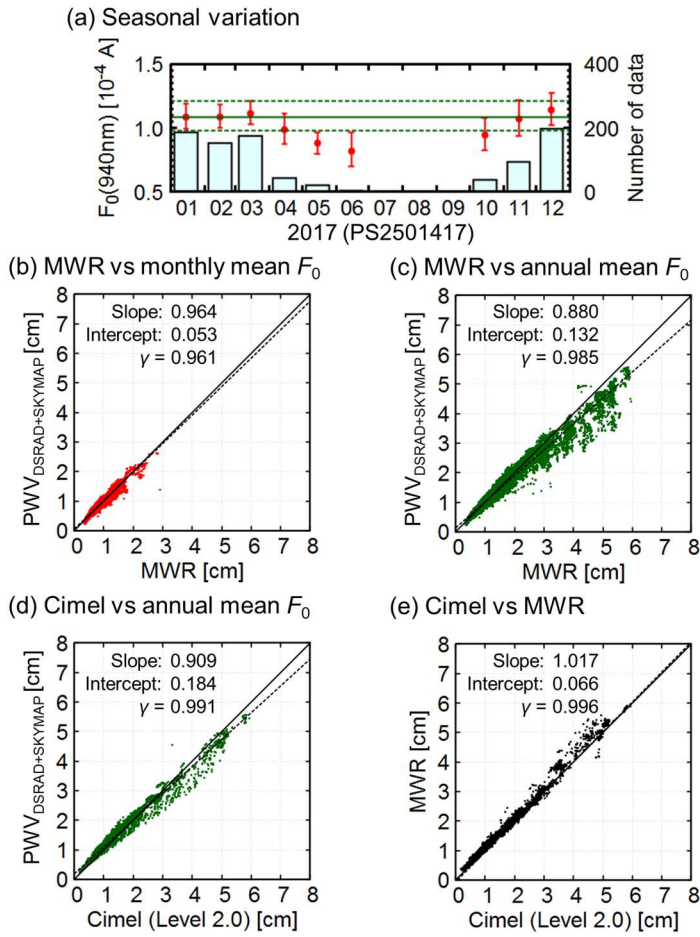


Figure 1714: Application of our methods to observational data from Chiba in 2017. (a) Seasonal variation in the calibration constant of the water vapor channel (red circles and error bars are monthly means and standard deviations, respectively; green solid and dotted lines are annual means and standard deviations, respectively; boxes indicate the number of data points). (b), (c) Comparison of PWV between the MWR and the sky-radiometer with (b) the monthly mean F_0 , and (c) the annual mean F_0 . (d) Comparison of PWV between the Cimel level 2.0 data and the sky-radiometer with annual mean F_0 . (e) Comparison of PWV between the Cimel level 2.0 data and the MWR.

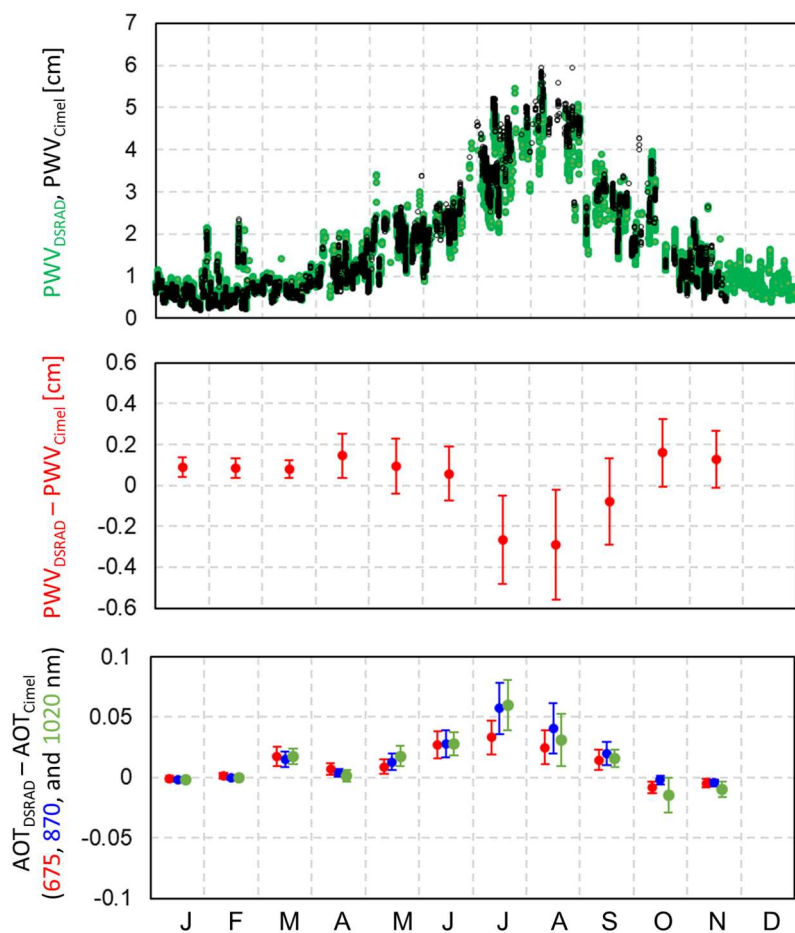


Figure 18: The top row shows the time series of PWV in 2017 at Chiba (green and black circles are PWV_{DSRAD+SKYMAP} and PWV_{Cimel}, respectively). The middle row is the difference between PWV_{DSRAD+SKYMAP} and PWV_{Cimel}. The bottom row is the difference in aerosol optical thicknesses at 675 nm (red), 870 nm (blue), and 1020 nm (green) between the DSRAD algorithm and the AERONET retrieval results. Circles and error bars in the middle and bottom rows are means and standard deviations, respectively.

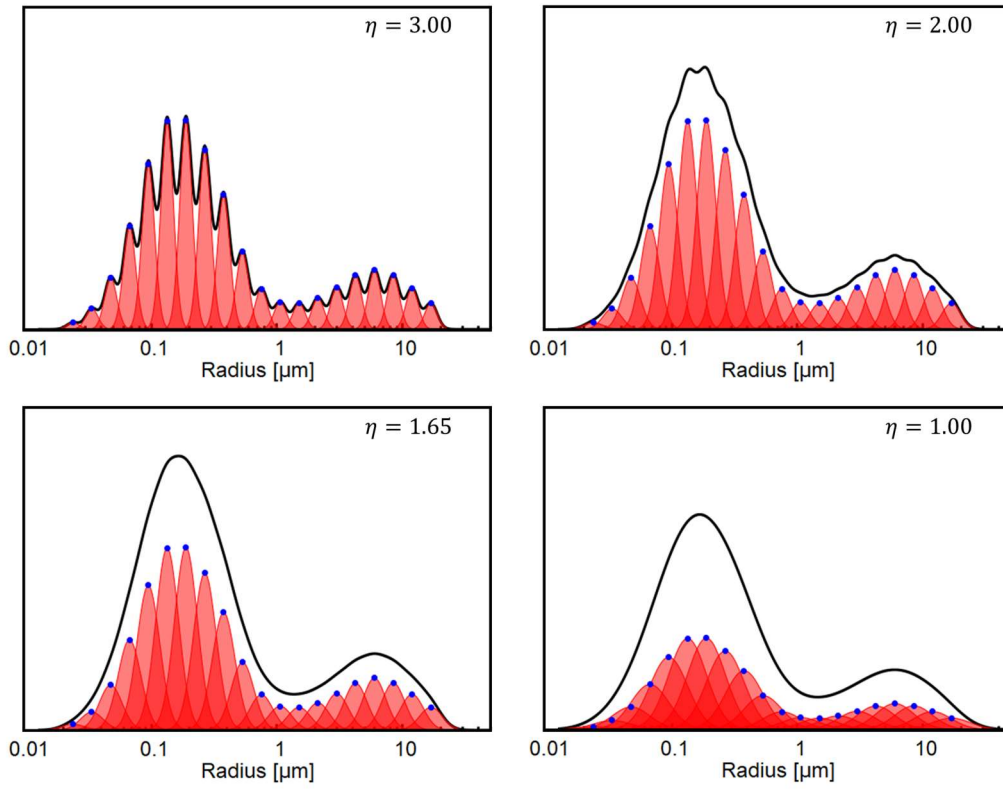


Figure A1: Relationship between the volume size distribution and η . The black line is the volume size distribution, which is computed by the integration of 20-modal lognormal distribution functions (red lines). Blue circles are the peak volume of lognormal size distribution.

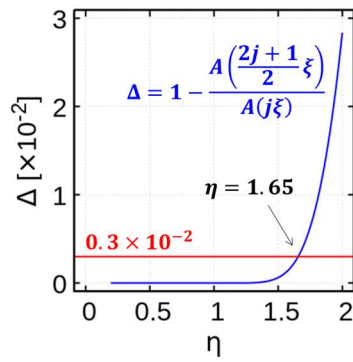


Figure A2: Relationship between the parameter η and the difference Δ .

Table 1: Sky-radiometer specifications. Each sky-radiometer is equipped with a filter indicated by a circle. “Standard” is the standard specification of sky-radiometer models POM-01 and POM-02.

<u>Wavelength</u> <u>[nm]</u>	<u>Strong gas</u> <u>absorption</u>	<u>Main target</u> <u>substance</u>	<u>POM-01</u> <u>Standard</u>	<u>POM-02</u> <u>Standard</u>	<u>POM-02</u> <u>PS1202091</u>	<u>POM-02</u> <u>PS2501417</u>
<u>315</u>	<u>O₃</u>	<u>Ozone</u>	<u>○</u>	<u>○</u>	<u>=</u>	<u>○</u>
<u>340</u>	<u>=</u>	<u>Aerosol</u>	<u>=</u>	<u>○</u>	<u>○</u>	<u>○</u>
<u>380</u>	<u>=</u>	<u>Aerosol</u>	<u>=</u>	<u>○</u>	<u>○</u>	<u>○</u>
<u>400</u>	<u>=</u>	<u>Aerosol</u>	<u>○</u>	<u>○</u>	<u>○</u>	<u>○</u>
<u>500</u>	<u>=</u>	<u>Aerosol</u>	<u>○</u>	<u>○</u>	<u>○</u>	<u>○</u>
<u>675</u>	<u>=</u>	<u>Aerosol</u>	<u>○</u>	<u>○</u>	<u>○</u>	<u>○</u>
<u>870</u>	<u>=</u>	<u>Aerosol</u>	<u>○</u>	<u>○</u>	<u>○</u>	<u>○</u>
<u>940</u>	<u>H₂O</u>	<u>Water vapor</u>	<u>○</u>	<u>○</u>	<u>○</u>	<u>○</u>
<u>1020</u>	<u>=</u>	<u>Aerosol</u>	<u>○</u>	<u>○</u>	<u>○</u>	<u>○</u>
<u>1225</u>	<u>O₂, CO₂, H₂O</u>	<u>Cloud</u>	<u>=</u>	<u>=</u>	<u>○</u>	<u>=</u>
<u>1627</u>	<u>CH₄, CO₂</u>	<u>Cloud</u>	<u>=</u>	<u>○</u>	<u>○</u>	<u>○</u>
<u>2200</u>	<u>CH₄, H₂O</u>	<u>Cloud</u>	<u>=</u>	<u>○</u>	<u>○</u>	<u>○</u>

995 **Table 21: Microphysical and optical properties and vertical profiles of aerosol used in sensitivity tests.**

Aerosol	Components	Particle shape	Size distribution		Refractive index at 940 nm		Relative weight in total optical thickness at 500 nm	Vertical profile
			Mode radius (μm)	Mode width	Real	Imaginary		
Continental average	Water-soluble	Sphere	0.18	0.81	1.43	0.0074	0.90	$\exp(-z/H)$, $H = 8$ km
	Soot	Sphere	0.05	0.69	1.75	0.44	0.07	$\exp(-z/H)$, $H = 4$ km
	Insoluble	Spheroid	5.98	0.92	1.52	0.008	0.03	$\exp(-z/H)$, $H = 2$ km
Transported dust	Dust	Spheroid	3.23	0.79	1.53	0.004	0.25	$\frac{1}{\sqrt{2\pi}\sigma} \exp\left(-\frac{(z-z_c)^2}{2\sigma^2}\right)$, $z_c = 3.5$ km $\sigma = 0.4$ km
	Water-soluble	Sphere	0.18	0.81	1.43	0.0074	0.67	$\exp(-z/H)$, $H = 8$ km
	Soot	Sphere	0.05	0.69	1.75	0.44	0.05	$\exp(-z/H)$, $H = 4$ km
	Insoluble	Spheroid	5.98	0.92	1.52	0.008	0.03	$\exp(-z/H)$, $H = 2$ km

<u>Aerosol</u>	<u>Componen</u> <u>ts</u>	<u>Particle</u> <u>shape</u>	<u>Size</u> <u>distribution</u>		<u>Refractive</u> <u>index at 940 nm</u>		<u>Relative</u> <u>weight</u> <u>in total</u> <u>optical</u> <u>thicknes</u> <u>s at</u> <u>500 nm</u>	<u>Vertical profile</u>
			<u>Mod</u> <u>e</u> <u>radiu</u> <u>s</u> <u>(μm)</u>	<u>Mod</u> <u>e</u> <u>widt</u> <u>h</u>	<u>Rea</u> <u>l</u>	<u>Imaginar</u> <u>y</u>		
<u>Continent</u> <u>al average</u>	<u>Water-</u> <u>soluble</u>	<u>Sphere</u>	<u>0.18</u>	<u>0.81</u>	<u>1.4</u> <u>3</u>	<u>0.0074</u>	<u>0.90</u>	$\exp(-z/H)$, $H = 8$ km
	<u>Soot</u>	<u>Sphere</u>	<u>0.05</u>	<u>0.69</u>	<u>1.7</u> <u>5</u>	<u>0.44</u>	<u>0.07</u>	$\exp(-z/H)$, $H = 4$ km

	<u>Insoluble</u>	<u>Spheroid</u>	<u>5.98</u>	<u>0.92</u>	<u>1.5</u>	<u>0.008</u>	<u>0.03</u>	$\exp(-z/H)$
		<u>d</u>			<u>2</u>			$H = 2 \text{ km}$
<u>Transported dust</u>	<u>Dust</u>	<u>Spheroid</u>	<u>3.23</u>	<u>0.79</u>	<u>1.5</u>	<u>0.004</u>	<u>0.25</u>	$\frac{1}{\sqrt{2\pi}\sigma} \exp\left(-\frac{(z-z_c)^2}{2\sigma^2}\right)$
		<u>d</u>			<u>3</u>			$z_c = 3.5 \text{ km}$
								$\sigma = 0.4 \text{ km}$
	<u>Water-soluble</u>	<u>Sphere</u>	<u>0.18</u>	<u>0.81</u>	<u>1.4</u>	<u>0.0074</u>	<u>0.67</u>	$\exp(-z/H)$
					<u>3</u>			$H = 8 \text{ km}$
	<u>Soot</u>	<u>Sphere</u>	<u>0.05</u>	<u>0.69</u>	<u>1.7</u>	<u>0.44</u>	<u>0.05</u>	$\exp(-z/H)$
					<u>5</u>			$H = 4 \text{ km}$
	<u>Insoluble</u>	<u>Spheroid</u>	<u>5.98</u>	<u>0.92</u>	<u>1.5</u>	<u>0.008</u>	<u>0.03</u>	$\exp(-z/H)$
		<u>d</u>			<u>2</u>			$H = 2 \text{ km}$

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Table 32: Validation of the SCAD method by visual observation from 2013 to 2014 in Tsukuba.

Visual observation	Sky-radiometer measuring plane	
Cloud cover	Clear sky	Cloud affected
Clear, less than 1	463 (83.4%)*	68 (9.3%)
Cloud affected, more than 2	92 (16.6%)*	663(90.7%)*

<u>Visual observation</u>	<u>Sky-radiometer measuring plane</u>	
<u>Cloud cover</u>	<u>Best condition</u>	<u>Poor condition</u>
<u>Clear, less than 20%</u>	<u>463 (83.4%)*</u>	<u>68 (8.7%)</u>
<u>Cloud affected, more than 20%</u>	<u>92 (16.6%)</u>	<u>714(91.3%)*</u>

*Obviously correct determination.

~~†Obviously incorrect determination.~~

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Table 43: References and methodologies of the DSRAD algorithm.

DSRAD	
Solar coordinates	Nagasawa (1999)
Refraction correction	Nagasawa (1999)
Sun-Earth distance	Nagasawa (1999)
Optical mass	Gueymard (2001)
Rayleigh scattering	Fröhlich and Shaw (1980); Young(1981)
Ozone absorption	Correlated k -distribution
Water vapor absorption	Correlated k -distribution
Filter response function	Stepwise function
Retrieved the PWV	Newton-Raphson method

DSRAD

<u>Solar coordinates</u>	<u>Nagasawa (1999)</u>
<u>Refraction correction</u>	<u>Nagasawa (1999)</u>
<u>Sun-Earth distance</u>	<u>Nagasawa (1999)</u>
<u>Optical mass</u>	<u>Gueymard (2001)</u>
<u>Rayleigh scattering</u>	<u>Fröhlich and Shaw (1980); Young(1981)</u>
<u>Ozone absorption</u>	<u>Sekiguchi and Nakajima (2008)</u>
<u>Water vapor absorption</u>	<u>Sekiguchi and Nakajima (2008)</u>
<u>Filter response function</u>	<u>Stepwise function</u>
<u>Retrieval of PWV</u>	<u>Newton-Raphson method</u>

Table 5: Comparison of PWV between DSRAD and other instruments.

		<u>Slope</u> C_1	<u>Intercept</u> C_2 [cm]	γ	<u>Bias</u> [cm]	<u>RMSE</u> [cm]
<u>PS1202091 at Tsukuba, Japan</u>						
<u>Monthly mean</u> F_0	<u>vs GNSS/GPS receiver</u> <u>(2013)</u>	<u>0.956</u>	<u>0.079</u>	<u>0.938</u>	<u>-0.049</u>	<u>0.138</u>
	<u>vs GNSS/GPS receiver</u> <u>(2014)</u>	<u>0.937</u>	<u>0.161</u>	<u>0.970</u>	<u>-0.110</u>	<u>0.170</u>
<u>Annual mean</u> F_0	<u>vs GNSS/GPS receiver</u> <u>(2013)</u>	<u>0.919</u>	<u>0.173</u>	<u>0.987</u>	<u>-0.061</u>	<u>0.226</u>
	<u>vs GNSS/GPS receiver</u> <u>(2014)</u>	<u>0.934</u>	<u>0.178</u>	<u>0.987</u>	<u>-0.089</u>	<u>0.223</u>
<u>PS2501417 at Chiba, Japan</u>						
<u>Monthly mean</u> F_0	<u>vs MWR (2017)</u>	<u>0.964</u>	<u>0.053</u>	<u>0.961</u>	<u>-0.027</u>	<u>0.091</u>
	<u>vs AERONET (2017)</u>	<u>0.987</u>	<u>0.107</u>	<u>0.976</u>	<u>0.098</u>	<u>0.122</u>
<u>Annual mean</u> F_0	<u>vs MWR (2017)</u>	<u>0.880</u>	<u>0.132</u>	<u>0.985</u>	<u>0.042</u>	<u>0.231</u>
	<u>vs AERONET (2017)</u>	<u>0.909</u>	<u>0.184</u>	<u>0.991</u>	<u>0.055</u>	<u>0.186</u>

$$C_1, C_2: PWV_{DSRAD} = C_1 \times PWV_{Other} + C_2$$

$$\text{Bias: } PWV_{DSRAD} - PWV_{Other}$$

Table 6: Difference in PWV between DSRAD with the annual mean calibration constants and other instruments.

	<u>PWV_{Other}</u>				
	<u>0 – 1 cm</u>	<u>1 – 2 cm</u>	<u>2 – 3 cm</u>	<u>3 – 4 cm</u>	<u>> 4 cm</u>
	<u>Bias [cm]</u>	<u>Bias [cm]</u>	<u>Bias [cm]</u>	<u>Bias [cm]</u>	<u>Bias [cm]</u>
	<u>(RMSE [cm])</u>	<u>(RMSE [cm])</u>	<u>(RMSE [cm])</u>	<u>(RMSE [cm])</u>	<u>(RMSE [cm])</u>
<u>PS1202091 at Tsukuba, Japan</u>					
<u>vs GNSS/GPS receiver (2013)</u>	<u>0.083</u> <u>(0.124)</u>	<u>0.160</u> <u>(0.211)</u>	<u>0.084</u> <u>(0.236)</u>	<u>-0.098</u> <u>(0.326)</u>	<u>-0.339</u> <u>(0.537)</u>
<u>vs GNSS/GPS receiver (2014)</u>	<u>0.110</u> <u>(0.142)</u>	<u>0.163</u> <u>(0.221)</u>	<u>0.107</u> <u>(0.251)</u>	<u>-0.055</u> <u>(0.353)</u>	<u>-0.239</u> <u>(0.492)</u>
<u>PS2501417 at Chiba, Japan</u>					
<u>vs MWR (2017)</u>	<u>0.017</u> <u>(0.066)</u>	<u>0.024</u> <u>(0.153)</u>	<u>-0.041</u> <u>(0.212)</u>	<u>-0.356</u> <u>(0.465)</u>	<u>-0.594</u> <u>(0.722)</u>
<u>vs AERONET (2017)</u>	<u>0.088</u> <u>(0.105)</u>	<u>0.118</u> <u>(0.192)</u>	<u>0.017</u> <u>(0.223)</u>	<u>-0.214</u> <u>(0.386)</u>	<u>-0.264</u> <u>(0.306)</u>
<u>Bias: PWV_{DSRAD} – PWV_{Other}</u>					

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