Assessment of TOMS UV bias due to absorbing aerosols

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[1] We compared NASA Total Ozone Mapping Spectrometer (TOMS) overpass irradiances against ground-based Brewer measurements at Ispra, Italy, and Thessaloniki, Greece, with the main objective of evaluating the effect of absorbing aerosols on TOMS UV bias using direct measurements of aerosol optical properties. Dependence of the TOMS/Brewer bias on aerosol absorption optical depth is significant. Our study demonstrates that the bias between TOMS and measured Brewer 324 nm irradiances increases with increasing aerosol absorption optical depth τ_{abs}. The correlation coefficients between the ratio TOMS/Brewer and τ_{abs} at Ispra and Thessaloniki are $\sim$0.7 or more and $\sim$0.8 or more, respectively, depending on the range of solar zenith angle values selected for the analysis. We found that the correlation with single-scattering albedo ω or aerosol optical depth is significantly smaller than with τ_{abs}. If measurements or climatology of τ_{abs} are available in the UVB range, the TOMS UV product can be postcorrected for absorbing aerosols using similar site-dependent regressions as established in our study. If no correction is applied, the mean positive biases between TOMS and Brewer 324 nm irradiances in Thessaloniki and Ispra are 19.2% and 29.7%, respectively, while the standard deviations are 8.9% and 16.5%. Depending on the level of correction, the mean positive bias can be essentially removed, and the standard deviations can be reduced down to $\sim$6% and $\sim$10%, respectively.

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1. Introduction

[2] Since the network of ground-based UV measurements will inevitably remain sparse, satellite-based UV methods offer a complementary approach to better document the geographical distribution of surface UV irradiance. Surface UV estimates based on Total Ozone Mapping Spectrometer (TOMS) satellite data have been used extensively in the last decade to establish global UV climatologies and to examine possible long-term changes in surface UV [*Herman et al.*, 1999; *Krotkov et al.*, 2002].

[3] The differences between ground-based UV measurements and satellite-derived estimates result from intrinsic differences in the two approaches. Ground instruments measure surface irradiance directly and are therefore subject only to instrumental errors (random and systematic) [*Bernhard and Seckmeyer*, 1999]. In contrast, since satellite UV maps are derived composite products, they are affected by both measurement errors (calibration, noise) and by modeling uncertainties. TOMS UV maps are calculated from satellite measured ozone, aerosol, and UV reflectivity data [*Krotkov et al.*, 1998; *Herman et al.*, 1999; *Krotkov et al.*, 2001]. Typically, uncertainties in TOMS total ozone and reflectivity data are small [*Herman et al.*, 1999], so that the main uncertainties in TOMS UV data arise from other uncertainties in the radiative transfer model input assumptions [*Krotkov et al.*, 2002]. The difference in spatial resolution and geometry between satellite and ground UV data is another important consideration in analyzing the differences; since the satellite field of view (FOV) is quite large ($\sim$100 km for TOMS), a single ground instrument measurement typically does not represent the whole satellite FOV area. Short-term variability in cloud field structures also has different effects on ground-based and satellite measurements [*Krotkov et al.*, 2002]. Nevertheless, the validation of satellite-derived UV products using ground-based measurements is an essential task in order to assess the accuracy of the products. Several validation studies [e.g., *Herman et al.*, 1999; *Kalliskota et al.*, 2000;

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McKenzie et al., 2001; *Arola et al.*, 2002; *Chubarova et al.*, 2002; *Fioletov et al.*, 2002, 2004] have revealed a positive (TOMS UV being higher) summertime bias in many locations, while a negative bias has been found in winter [*Kalliskota et al.*, 2000; *Fioletov et al.*, 2004]. The bias can be seen both at ozone-absorbing and nonabsorbing wavelengths, as well as in clear-sky conditions. This suggests the difference is not related to ozone absorption or cloud effects. It has also been observed that TOMS UV agrees much better with ground-based measurements at pristine sites (Lauder, in New Zealand [*McKenzie et al.*, 2001], or the Canadian west coast [*Fioletov et al.*, 2002]) than at more polluted European and North American sites. It has therefore been suggested that at least part of the bias could be attributed to boundary layer aerosol absorption that is not accounted for in the current TOMS UV algorithm [*Krotkov et al.*, 2002, 2005a]. *Krotkov et al.* [1998] have estimated the amount of aerosol absorption required to reconcile the bias between the Brewer-measured and TOMS-estimated UV irradiances. Using the Brewer-measured extinction optical depth, they estimated an average aerosol single-scattering albedo ω of 0.95 at 325 nm on clear summer days in Toronto Canada.

[4] *Torres et al.* [2005] first derived ω in the UV using TOMS satellite measurements for the Southern African Regional Science Initiative (SAFARI) 2000 campaign in Africa. However, these indirect estimates have not been verified by colocated direct measurements of aerosol absorption properties in the UV.

[5] In fact, direct measurements of aerosol absorption present a great challenge even for ground-based remote sensing techniques. So far, direct ω retrievals in the UV have been reported using the Multi-Filter Rotating Shadowband Radiometer (MFRSR) measurements at Black Mountain, North Carolina [*Petters et al.*, 2003], and the combined UV-MFRSR and Aerosol Robotic Network (AERONET)/Cimel measurements at the NASA GSFC site at Greenbelt, Maryland [*Krotkov et al.*, 2005b]. At the GSFC site, the summer case average ω was slightly less in the UV ($\omega_{368} > 0.93$) than in the visible ($\omega_{440} > 0.95$). However, the spectral differences were within the overlap of the error bars for both techniques [*Krotkov et al.*, 2005b]. Since this measurement was only for urban aerosols, it is important to conduct direct measurements of aerosol absorption using different techniques and in different locations; these include especially those affected by desert dust and cities having urban pollution, where stronger UV aerosol absorption is expected.

[6] Our main objective in this paper was to combine ground measurements of aerosol UV absorption and global UV irradiance with TOMS satellite estimations of UV irradiance, in an attempt to explain the observed differences between them. We have compared TOMS overpass irradiances with the corresponding Brewer measurements (averaged within a 2-hour window around the overpass time) at Ispra, Italy, and Thessaloniki, Greece, with the emphasis being placed on an assessment of the effect of absorbing aerosols on surface UV irradiance and TOMS UV bias. The values of single-scattering albedo ω were estimated from the Brewer global irradiance measurements with the aid of radiative transfer modeling.

2. Data and Methodology

[7] Irradiance spectra from Brewer spectrometers at Ispra, Italy, and Thessaloniki, Greece, were used in this study. Ground-based spectral data were averaged between 1100 and 1300 local solar time, to reduce the influence of temporal changes in cloud cover. Data from both sites have been corrected for the nonideal angular response of their input optics (cosine correction). The main details of the cosine correction of the instrument at Thessaloniki are described by *Bais et al.* [1998], and the methodology of cosine correction at Ispra is essentially the same. The TOMS spectral UV time series were reprocessed to use the average solar zenith angle of the ground measurements within the selected time interval.

[8] The new TOMS UV product includes surface irradiance estimates at the following wavelengths: 305, 310, 324, and 380 nm ± 0.25 nm with a triangular slit function to approximately match the Brewer spectral resolution [*Krotkov et al.*, 2002, 2005a]. We selected the 324 nm irradiances for the comparisons, because this is the highest common wavelength that both spectroradiometers are able to measure and is least dependent on ozone variations.

2.1. TOMS UV Estimates

[9] The TOMS UV algorithm for estimating surface UV irradiance from satellite measurements proceeds in two steps: First, the clear-sky radiation, E_{clear} is calculated, and second, E_{clear} is multiplied with the estimated cloud transmission factor C_T [*Herman et al.*, 1999; *Krotkov et al.*, 1998, 2001]. As previously described in the literature, calculation of E_{clear} in the UV spectral range is obtained from the satellite-derived spectral extraterrestrial solar irradiance (ATLAS-3 SUSIM) and TOMS measurements of total column ozone and scene reflectivity. Calculation of E_{clear} in the UV range from satellite-derived spectral extraterrestrial solar irradiance and TOMS measurements of total column ozone, aerosols and surface reflectivity is described in detail by *Krotkov et al.* [1998] and *Herman et al.* [1999].

[10] In the absence of snow, clouds and aerosols, the effects of molecular (Rayleigh) scattering, ozone absorption, solar zenith angle, and altitude are well-understood problems. However, the presence of aerosols, clouds and snow in the TOMS FOV requires additional corrections. The type of correction (the specific C_T algorithm) selected is based on two threshold values of the aerosol index (AI) (calculated from 331 nm and 360 nm radiances) and the Lambertian Equivalent Reflectivity (LER) at 360 nm. For TOMS, AI is defined as

$$\mathrm{AI} = -100\left[\log 10\left(\frac{I_{331}}{I_{360}}\right)_{\text{measured}} - \log 10\left(\frac{I_{331}}{I_{360}}\right)_{\text{calculated}}\right], \quad (1)$$

where I_{331} and I_{360} are TOMS measured and calculated reflectances at 331 nm and 360 nm (using LER derived at 360 nm).

[11] Currently, if the AI is larger than 0.5 and the reflectivity at 360 nm is less than 0.15 then absorbing aerosols are assumed and an AI correction is applied. Otherwise, the cloud C_T model is assumed; the algorithm does not, therefore, distinguish between thin clouds and aerosols [*Krotkov et al.*, 2005a]. This causes a typical C_T error of 2% for nonabsorbing sulphate or sea salt aerosols with an aerosol optical depth of 0.2. However, cloud CT errors become large (up to +30%) in case of urban and carbonaceous absorbing aerosols. Since such aerosols are located in the boundary layer, they typically produce AI < 0.5 and cannot be separated from nonabsorbing aerosols using AI data [*Krotkov et al.*, 2002].

[12] The surface albedo and snow effects are estimated using the TOMS monthly Minimum Lambertian Effective Surface Reflectivity (MLER) global database [*Herman and Celarier*, 1997], as described by *Krotkov et al.* [1998] and *Krotkov et al.* [2002].

2.2. Spectral Irradiance Measurements

[13] The MK II Brewer at Ispra is a single-pass monochromator with a spectral range of 290–325 nm, measuring with a step of 0.5 nm and FWHM of 0.55 nm [*Gröbner and Meleti*, 2004]. The Ispra Brewer makes an up scan followed by a down scan, and the mean of the two scans is used for a representative spectrum at one fixed time. The whole up and down measurement lasts for 5 min.

[14] The Brewer spectroradiometer situated in the Laboratory of Atmospheric Physics in Thessaloniki, Greece, is a double monochromator [*Bais et al.*, 1996] operating in the spectral range 287.5–366.0 nm in steps of 0.5 nm, with a spectral resolution of 0.55 nm (FWHM). The time for completing one spectrum is ~6–7 min. Both Brewers have a 35-mm-diameter Teflon diffuser, which is protected by a weatherproof quartz dome.

2.3. Aerosol Optical Depth Retrieval

[15] The aerosol optical depth (AOD) was determined by comparing the measured direct irradiance spectra to the direct irradiance at the surface. This surface irradiance was calculated from the transfer of the validated extraterrestrial irradiance (ATLAS 3 solar spectrum) through the atmosphere, accounting for molecular absorption (mainly ozone and SO_2) and Rayleigh scattering [*Gröbner and Kerr*, 2001; *Marenco et al.*, 1997]. Essentially the only difference in AOD retrievals between these sites is that the Brewer at Ispra uses measurements in the UVB (5 wavelengths using the direct Sun Brewer routine), while the Thessaloniki Brewer uses spectral (300–365 nm with a step of 0.5 nm) measurements.

[16] For the absolute calibration of the direct irradiance, it is compared with the difference between global and diffuse irradiance (obtained with the use of a shadow disk), which are measured nearly simultaneously [*Bais*, 1997]. The procedures to retrieve the AOD from the Brewer measurements at Ispra and Thessaloniki are described in detail by *Gröbner and Meleti* [2004] and *Kazadzis et al.* [2005], respectively.

[17] The mean annual variation of AOD at 340 nm over Thessaloniki ranges from 0.3 in December to 0.7 in August. The observed seasonality can be attributed to various local parameters, such as the enhanced evaporation and high temperatures in the summer that increase the turbidity, the absence of any significant wet removal of aerosols, and the transboundary transportation of particles. The latter is confirmed by the consistency of the back trajectories [*Formenti et al.*, 2001; *Gerasopoulos et al.*, 2003]. Additionally, northern winds are dominant during the wintertime, which leads to a "cleaning" of the atmosphere and thus to lower AOD values.

[18] At Ispra the AOD values at 320 nm are lowest in the winter months, with a mean AOD of 0.3. In spring and summer the mean AOD ranges from around 0.5 (March, April) to 0.65 in July [*Gröbner and Meleti*, 2004].

2.4. Single-Scattering Albedo ω Retrieval

[19] The retrieval of the single-scattering albedo ω employed the Brewer global irradiance measurements and radiative transfer (RT) modeling. We used the LibRadtran 0.99 package and the UVSPEC DISORT radiative transfer equation solver (code is freely available from http://www.libradtran.org).

[20] For the estimation of ω, we adopted a look-up table (LUT) approach, in which the values of global irradiance were tabulated as a function of the variable input parameters: SZA, ω, and AOD. The values of global irradiance at 320 nm were used to retrieve ω from the single Brewer at Ispra; the total ozone was therefore an additional input variable. However, since ω from the double Brewer at Thessaloniki was derived using the global irradiance at 340 nm, the effect of ozone was considered negligible.

[21] The slit function of the instrument was also incorporated in modeling the irradiances. Given that the other input variables are known, the LUT can be used to estimate a single unknown variable corresponding to each global measurement. The values for the other necessary input variables were assumed (an aerosol asymmetry factor of 0.7 and urban profiles of aerosols according to *Shettle* [1989]) or known (e.g., Sun-Earth distance and the altitude of the station). The asymmetry parameter (g) was not measured at UV wavelengths neither at Ispra, nor at Thessaloniki. However, typical uncertainty in g (~0.05) would result in less than 0.02 error in ω [*Krotkov et al.*, 2005b]. This is less than Brewer measurement errors in estimating ω (~0.05 due to calibrations and other uncertainties). Therefore we have not made an effort to constrain the g parameter from ancillary measurements and rather used generic value 0.7 at both places.

[22] Since the SZA, AOD and total ozone for each global measurement are known, the corresponding ω can be interpolated from the LUT of global irradiances tabulated as described above. We used third-order Lagrangian interpolation for this purpose. A total estimated uncertainty for single-scattering albedo estimates of 0.13 for an AOD of 0.2 and 0.05 for an AOD of 0.4 have been reported [*Bais et al.*, 2005].

[23] In our study, following the uncertainty analysis reported by *Bais et al.* [2005] and taking into account the mean AOD values for the both sites, the uncertainties of analyzed ω measurements are estimated being at the range of 0.04 and 0.09 for summer and winter, respectively. The mean uncertainty for the absolute single-scattering albedo

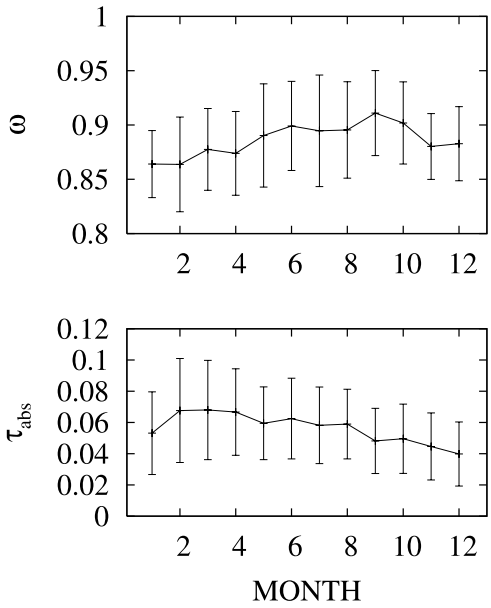

Figure 1. Mean monthly (top) ω and (bottom) τ_{abs} from Brewer data at 320 nm at Ispra and the associated standard deviation (1σ).

determination is on the order of 0.05 and 0.06 for Thessaloniki and Ispra, respectively.

3. Results and Discussion

[24] Absorbing aerosols in the boundary layer are capable of attenuating UV irradiance more strongly than nonabsorbing aerosols for the same aerosol optical depth [*Krotkov et al.*, 1998]. This effect would cause the cloud C_T correction to underestimate the actual attenuation of the surface UV irradiance. Moreover, since these aerosols also attenuate the outgoing shortwave radiation, the cloud C_T algorithm underestimates aerosol optical depth, amplifying the error further, causing an overestimation of UV irradiance in the summertime. Because pollution aerosols are typically located in the boundary layer, they tend to produce a negative AI that makes it impossible to distinguish from the nonabsorbing aerosols and thin clouds using AI data alone. To quantify the C_T error due to aerosol absorption, one needs actual measurements of aerosol absorption optical depth, τ_{abs} = AOD · (1 − ω) [*Krotkov et al.*, 2005a].

[25] Figure 1 shows the mean monthly values of ω and τ_{abs} derived from Brewer global irradiance measurements at 320 nm at Ispra and the corresponding standard deviation (1σ), calculated from the daily mean values for every month of the period 1992–2002. There is a significant variability in ω within each month, while there also exists a seasonal variation in the monthly means, with the lowest values in the winter and increasing values from winter until late summer. A marked variability can be also observed in τ_{abs} with maximum in late winter–early spring and a progressive minimum toward late fall. This differs from measurements in Greenbelt, Maryland, where a pronounced maximum (τ_{abs} = 0.08 at 325 nm) was measured in summer hazy conditions while τ_{abs} < 0.02 during cold periods [*Krotkov et al.*, 2005b].

[26] Aerosol absorption retrievals are performed assuming zero NO_2 amounts. However, recently, *Krotkov et al.* [2005c] demonstrated that NO_2 correction is required for accurate τ_{abs} retrievals, particularly for low-AOD cases.

[27] In Figure 1 and in the following analysis, we have included only the ω estimates for AOD larger than 0.3 in order to reduce the uncertainty. In this case, the estimated uncertainty for ω is always less than 0.1 and ∼0.02 for τ_{abs}.

[28] It is difficult to validate the single-scattering albedo data from the Brewer measurements directly against the independent AERONET Sun photometer ω inversions, since the lowest wavelength available from the AERONET ω data at Ispra is 441 nm. However, the monthly values of ω and τ_{abs} from AERONET 441 nm data agree with the Brewer data within the stated uncertainties [*Holben et al.*, 1998; *Dubovik et al.*, 2000].

[29] Figure 2 shows the ratio of TOMS and Brewer 324 nm irradiances at Ispra plotted against the aerosol absorption optical depth τ_{abs}, as defined above. From the entire period of 1992–2002, all the possible clear-sky data are included. Brewer measurements are averaged within a 2-hour window around the overpass time of TOMS measurement; that is, local noon irradiances are compared. We estimated the regression equations for 10° SZA bins, and in Figure 2, three ranges are shown as an example: 20°–30°, 40°–50°, and 60°–70°.

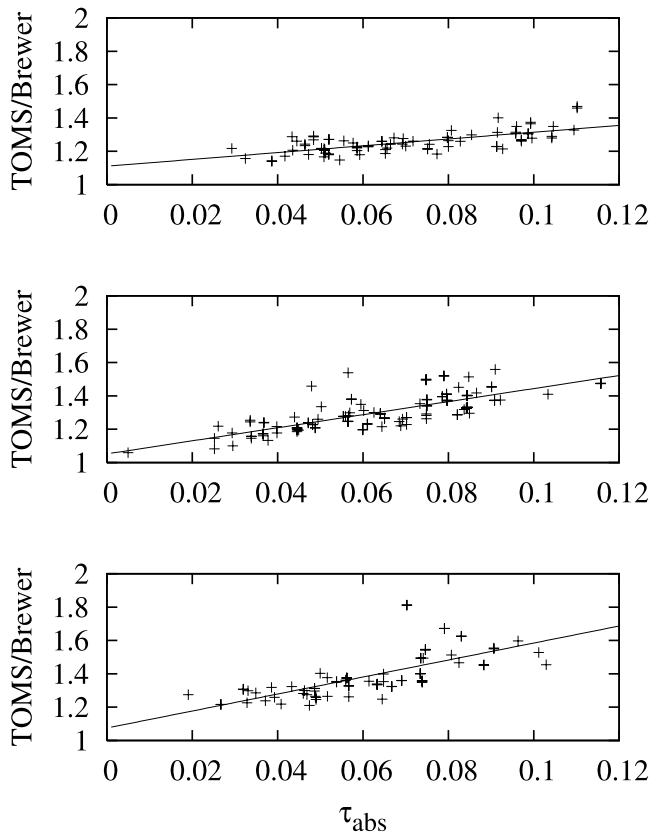

Figure 2. Ratio of TOMS to Brewer irradiances at 324 nm against aerosol absorption optical depth at Ispra. Estimates of τ_{abs} are derived from 320 nm irradiances. Three SZA ranges are shown: (top) 20°–30°, (middle) 40°–50°, and (bottom) 60°–70°.

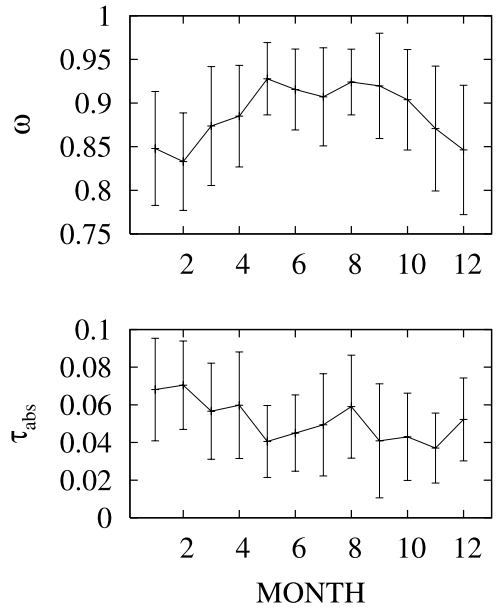

Figure 3. Mean monthly (top) ω and (bottom) τ_{abs} from Brewer data at 340 nm at Thessaloniki and the associated standard deviation (1σ).

[30] It is apparent that the TOMS bias at 324 nm is correlated with the aerosol absorption. The correlation coefficients of the fit are 0.69, 0.77, and 0.71 for the 20°–30°, 40°–50°, and 60°–70° bins, respectively. The bias increases with increasing aerosol absorption τ_{abs}, as one would assume, since the TOMS UV algorithm does not properly account for the absorbing aerosols in the boundary layer. The cases of AI greater than 0.5 and less than 0.5 would indicate dust aerosols and pollution aerosols, respectively. We included only the cases AI less than 0.5; that is, there is no correction applied for the absorbing aerosols in TOMS UV data. If the cases of TOMS greater than 0.5 were included (and thus AI correction was applied), these TOMS data would have partly accounted for absorbing aerosols, while τ_{abs} derived from the Brewer data accounts for the total column. Therefore, for consistency, we excluded the cases of AI greater than 0.5.

[31] We want to reemphasize here that τ_{abs} is the most appropriate quantity to evaluate the effect of absorbing aerosols on the TOMS/Brewer bias. The correlation of 0.18 between TOMS/Brewer and ω is much lower than that for τ_{abs}; similarly, the correlation is only 0.33 against the extinction aerosol optical depth, since TOMS UV algorithm mostly corrects for aerosol back scattering using thin cloud model.

[32] If the linear least squares fits shown in Figure 2 are extrapolated to zero τ_{abs}, one could estimate the remaining bias under aerosol-free conditions. Our results do suggest that there still remains a positive bias of ∼10%, not explained by the effect of absorbing aerosols. This remaining positive bias can be due to several sources of uncertainty both in satellite- and ground-based measurements. We would therefore rather emphasize that the more interesting quantity obtained in our study, focusing on the effect of absorbing aerosols, is the slope of the fit. This indicates the manner in which the bias increases with an increasing amount of aerosol absorption. The slopes of the fit in Figure 2 are 2.0, 3.9, and 5.1, while the slopes for the SZA bins of 30°–40° and 50°–60° are 2.8 and 3.4, respectively.

[33] Assuming relative standard uncertainties of 5% and 10% for Brewer and TOMS irradiance values (error in TOMS/Brewer ≈ 0.11) and 0.02 for τ_{abs} (error in $\tau_{abs} \approx \delta\omega \cdot AOD$), then the estimated uncertainties for the regression coefficients are ∼0.05 for the intercept and ∼0.8 for the slope estimate. It is stressed that these estimates are based on the estimated total uncertainty. However, for the purpose of our study, the systematic errors in τ_{abs} are not actually very critical. The systematic underestimation or overestimation of Brewer global irradiance would result in underestimated or overestimated ω values. This systematic error cannot cause a clear impact on the correlation coefficients, while the slope in our regression analysis would be affected.

[34] Figure 3 shows the mean monthly values of ω and τ_{abs} derived from Brewer global irradiances at Thessaloniki and the corresponding 1σ range of the variability; that is, this plot for Thessaloniki corresponds to that shown for Ispra in Figure 1. It is apparent that there is a more distinct seasonal variability in ω as compared to the measurements at Ispra. However, in a rough sense, the AOD has a similar seasonal pattern, i.e., highest values (∼0.7) in the summer months and early autumn, decreasing in later autumn and

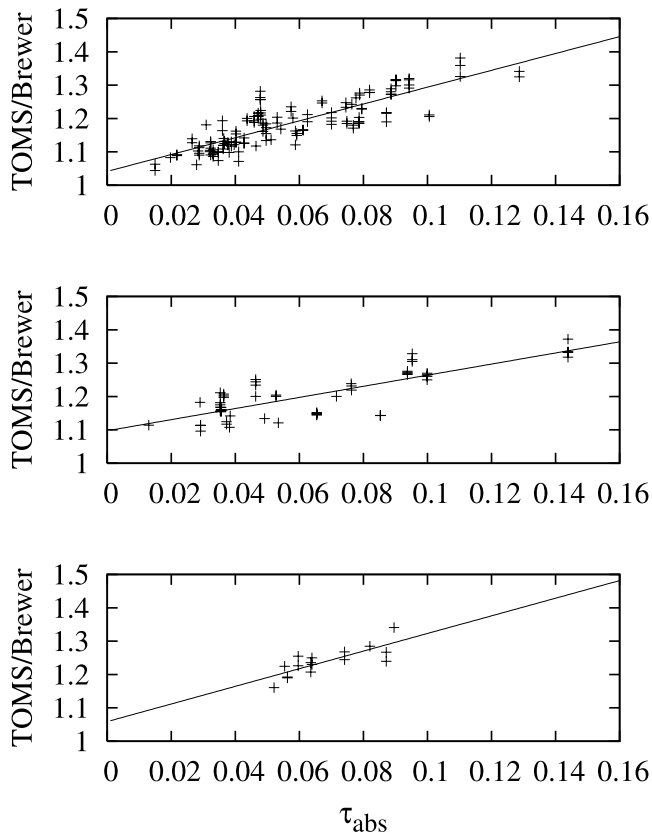

Figure 4. Ratio of TOMS to Brewer irradiances at 324 nm against aerosol absorption optical depth at Thessaloniki. Estimates of τ_{abs} are derived from 340 nm irradiances. Three SZA ranges are shown: (top) 20°–30°, (middle) 40°–50°, and (bottom) 60°–70°.

Table 1. Statistics of the Comparison Between Satellite- and Ground-Based Irradiances at Thessaloniki and Ispra[a]

	Thessaloniki		Ispra	
Type of Correction	Mean	SD	Mean	SD
CF = 1	19.2	8.9	29.7	16.5
CF = 1 + slope · $\overline{\tau_{abs}}$	5.1	8.1	6.8	13.2
CF = 1 + slope · τ_{abs}	4.8	6.8	5.6	11.5
CF = intercept + slope · τ_{abs}	0.0	6.0	−1.4	10.2

[a]SD, standard deviation. Differences are calculated as percentages, i.e., [(satellite − ground)/ground] × 100%. Results with different corrections for TOMS data are shown; $I_{corrected} = I/CF$. Here, slope and intercept are those estimated from linear regressions (examples in Figures 2 and 4), and $\overline{\tau_{abs}}$ is mean τ_{abs}.

winter (∼0.3) [*Kazadzis et al.*, 2005]. In contrast, there is not such a clear variability in τ_{abs}. Nevertheless, in winter, τ_{abs} at Thessaloniki exhibits essentially similar behavior to that at Ispra, moreover, showing a secondary maximum around August. The regional differences in τ_{abs} and its seasonal cycle need further investigation.

[35] Figure 4 shows the ratio of the TOMS and Brewer 324 nm irradiances at Thessaloniki plotted against aerosol absorption optical depth and for the same SZA bins as in Figure 2. The data set includes clear-sky cases for the period from 1998 to 2001. The correlation coefficients in the top, middle, and bottom plots are 0.85, 0.81, 0.79, respectively, indicating a somewhat better fit than for the Ispra data. One reason for the better correlation may be that ω and τ_{abs} in Thessaloniki were determined using 340 nm global irradiances, while 320 nm was used for the Ispra data, since the highest wavelength of the single Brewer is 325 nm. Retrieval of ω from 340 nm is likely more robust, since the signal of the global irradiance measurement is higher; moreover, ozone absorption does not introduce an additional source of uncertainty.

[36] The slopes of the fit for the SZA bins of 20–30, 30–40, 40–50, 50–60, and 60–70 are 2.5, 2.0, 1.7, 1.8, and 2.6, thus being smaller, on average, if compared to the case of Ispra. However, the estimated slopes (in Figures 2 and 4) agree reasonably well (within uncertainties) with the previous theoretical estimates [*Krotkov et al.*, 2005b].

[37] If measurements or climatology of ω and AOD were available in the UV range, the TOMS UV data could be postcorrected for absorbing aerosols using the regression between TOMS/Brewer and τ_{abs}, as established above. In Table 1 some statistics from the comparisons between the TOMS and Brewer data are shown if different correction factors are applied for the TOMS UV data,

$$I_{corrected} = \frac{I}{CF} \quad (2)$$

where I is the operational TOMS UV irradiance, $I_{corrected}$ is TOMS UV irradiance corrected for the effect of absorbing aerosols and CF is the correction factor. The forms of the suggested correction factors are listed in Table 1.

[38] If no correction is applied, the mean positive bias in Thessaloniki and Ispra (the comparisons in Figures 2 and 4, i.e., including all aerosol conditions) is 19.2% and 29.7%, respectively, while the standard deviations are 8.9% and 16.5%. If a simple correction is applied, which uses the regression slopes in Figures 2 and 4 together with the mean τ_{abs} of the entire data (essentially, this results in a constant correction factor), the mean positive bias in both sites is strongly reduced and the standard deviation is decreased as well. This correction is relevant if only climatology of τ_{abs} is available. In the third correction, in addition, the actual variability in τ_{abs} is taken into account. It can be seen that the mean positive bias is not affected, if compared to the previous correction, but the standard deviation is slightly improved. The fourth correction uses the actual intercept of the regressions and, not surprisingly, results in close to zero mean bias; moreover, the standard deviations are somewhat further improved. However, it has to be stressed that in this last case, additionally other sources of uncertainty and differences in satellite- and ground-based measurements, not only the effect of aerosol absorption, are taken into account.

4. Conclusions

[39] We compared TOMS overpass irradiances against Brewer measurements at Ispra, Italy, and Thessaloniki, Greece, with the main objective of evaluating the effect of absorbing aerosols with measurements of aerosol optical properties.

[40] The dependence of bias on aerosol absorption optical depth was apparent. The correlation coefficients between the ratio TOMS/Brewer and τ_{abs} at Ispra and Thessaloniki were ∼0.7 and ∼0.8 or more, respectively. The correlation with ω or AOD was significantly smaller than with τ_{abs}.

[41] If the measurements or climatology of τ_{abs} are available in the UVB range, TOMS UV product can be postcorrected for absorbing aerosols using the relationship between the ratio TOMS/Brewer and τ_{abs}, as shown in Figures 2 and 4. If no correction is applied, the mean positive biases in Thessaloniki and Ispra between TOMS and Brewer 324 nm irradiances are 19.2% and 29.7%, respectively, while the standard deviations are 8.9% and 16.5%. Depending on the level of correction, the mean positive bias can be essentially removed and the standard deviations reduced down to ∼6% and ∼10%, respectively. Our results also suggest that there is still a remaining positive bias not explained by the aerosol effect.

[42] **Acknowledgment.** We are grateful to two anonymous reviewers, whose very constructive comments greatly improved the manuscript.

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