Verification of the Atmospheric Infrared Sounder (AIRS) and the Microwave Limb Sounder (MLS) ozone algorithms based on retrieved daytime and night-time ozone

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Abstract. Ozone (O3) plays a significant role in weather and climate on regional to global spatial scales. Most studies on the variability in the total column of O3 (TCO) are typically carried out using daytime data. Based on knowledge of the chemistry and transport of O3, significant deviations between daytime and night-time O3 are only expected either in the planetary boundary layer (PBL) or high in the stratosphere or mesosphere, with little effect on the TCO. Hence, we expect the daytime and night-time TCO to be very similar. However, a detailed evaluation of satellite measurements of daytime and night-time TCO is still lacking, despite the existence of long-term records of both. Thus, comparing daytime and night-time TCOs provides a novel approach to verifying the retrieval algorithms of instruments such as the Atmospheric Infrared Sounder (AIRS) and the Microwave Limb Sounder (MLS). In addition, such a comparison also helps to assess the value of night-time TCO for scientific research. Applying this verification on the AIRS and the MLS data, we identified inconsistencies in observations of O3 from both satellite instruments. For AIRS, daytime–night-time differences were found over oceans resembling cloud cover patterns and over land, mostly over dry land areas, which is likely related to infrared surface emissivity. These differences point to issues with the representation of both processes in the AIRS retrieval algorithm. For MLS, a major issue was identified with the “ascending–descending” orbit flag, used to discriminate night-time and daytime MLS measurements. Disregarding this issue, MLS day–night differences were significantly smaller than AIRS day–night differences, providing additional support for the retrieval method origin of AIRS in stratospheric column ozone (SCO) day–night differences. MLS day–night differences are dominated by the upper-stratospheric and mesospheric diurnal O3 cycle. These results provide useful information for improving infrared O3 products.

1 Introduction

Atmospheric ozone (O3) is a key factor in the structure and dynamics of the Earth’s atmosphere (London, 1980). The 1987 Montreal Protocol on Substances that Deplete the Ozone Layer formally recognized the significant threat of chlorofluorocarbons and other O3-depleting substances (ODCs) to the O3 layer and marks the start of joint international efforts to reduce and ultimately phase-out the global production and consumption of ODCs (Velders et al., 2007). Indeed, concerns about changes in O3 due to catalytic chemistry involving anthropogenically produced chlorofluorocarbons has become an important topic for the scientific community, the general public, and governments (Fioletov et al., 2002).

In response to this concern and associated environmental policies, a large number of studies during the last 2 decades have focused on estimating long-term variations and trends in the stratospheric column of O3 (SCO). A summary of the state of the science is frequently reported in the quadrennial O3 assessment reports issued by the United Nations Envi-
The O$_3$ diurnal cycle depends on latitude, altitude, weather, and time. The variations in the diurnal cycle are less than 5% in the tropics and subtropics and increase to more than 15% in the upper stratosphere during the polar day near 70° N (Frith et al., 2020). Diurnal variations exist in atmospheric O$_3$ at certain altitudes. There are two distinct O$_3$ maxima in the typical vertical profile of the O$_3$ volume mixing ratio: one in the lower stratosphere and one in the mesosphere. The secondary maximum in the mesosphere is present during both day and night (Evans and Llewellyn, 1972; Hays and Roble, 1973). Chapman (1930) revealed the photochemical scheme in the mesosphere. The reactions of the Chapman cycle are important for us to understand diurnal O$_3$ variation.

\begin{align}
O_2 + hv & \rightarrow 2O(\lambda < 240 \text{ nm}), \\
O + O_2 + M & \rightarrow O_3 + M, \\
\text{(in which } M \text{ stands for an air molecule)}
\end{align}

\begin{align}
O_3 + O & \rightarrow 2O_2, \\
O_3 + hv & \rightarrow O_2 + O(\lambda < 1140 \text{ nm}).
\end{align}

In the daytime mesosphere, catalytic O$_3$ depletion by odd hydrogen has to be considered in addition to the Chapman cycle. The anti-correlation of O$_3$ and temperature is mainly due to the temperature dependence of the chemical rate coeffi-

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ficients (Craig and Ohring, 1958; Barnett et al., 1975). Huang et al. (2008, 1997) found midnight O$_3$ increases in the mesosphere, based on SABER and MLS data respectively. Zomerfelds et al. (1989) surmised that eddy transport may explain this increase, whereas Connor et al. (1994) stated that atmospheric tides are expected to cause systematic day–night variations.

During daytime, photolysis is the major loss process. The main night-time O$_3$ source in the mesosphere is atomic oxygen, whereas its sinks are atomic hydrogen and atomic oxygen (Smith and Marsh, 2005). In addition to O$_3$ chemical reactions with active hydrogen and molecular oxygen, the turbulent mass transport also plays an important role in the explanation of the secondary O$_3$ maximum (Sakazaki et al., 2013; Schanz et al., 2014).

Tropospheric O$_3$ is mainly produced during chemical reactions when mixtures of organic precursors (CH$_4$ and non-methane volatile organic carbon, NMVOC), CO, and nitrogen oxides (or NO$_x$) are exposed to the UV radiation in the troposphere (Simpson et al., 2014). At night, in the absence of sunlight, there is no O$_3$ production, but surface O$_3$ deposition and dark reactions transform the NO$_x$–VOC mixture and remove O$_3$. The dark chemistry affects O$_3$, and its key ingredients mainly depend on the reactions of two nocturnal nitrogen oxides, NO$_3$ (the nitrate radical) and N$_2$O$_5$ (dinitrogen pentoxide). NO$_3$ oxidizes VOCs at night, whereas the reaction of N$_2$O$_5$ with aerosol particles containing water removes NO$_3$. Both processes also remove O$_3$ at night (Brown et al., 2006).

The diurnal cycle of O$_3$ in the middle stratosphere had generally been considered small enough to be inconsequential, with known larger variations in the upper stratosphere and mesosphere (Prather, 1981; Pallister and Tuck, 1983). Later studies have highlighted observed and modelled peak-to-peak variations of the order of 5% or more in the middle stratosphere between 30 and 1 hPa (Sakazaki et al., 2013; Parrish et al., 2014; Schanz et al., 2014).

In terms of dynamics, vertical transport due to atmospheric tides is expected to contribute to diurnal O$_3$ variations at altitudes where background O$_3$ levels have a sharp vertical gradient (Sakazaki et al., 2013). The Brewer–Dobson circulation transports air upwards in the tropics, and polewards and downwards at high latitudes, with stronger transport towards the winter pole (Chipperfield et al., 2017).

The main objective of this paper is to analyse day–night differences in the AIRS TCO and the MLS SCO as well as in MLS upper atmospheric O$_3$ profiles. Section 2 discusses the data used. Section 3 presents results for AIRS, MLS, the comparison of AIRS with MLS, and an application of AIRS TCO data over the Pacific low-O$_3$ regions to highlight how day–night differences affect the use and interpretation of TCO data. Finally, Sect. 4 provides a brief summary and conclusions.

2 Data

2.1 AIRS total column of O$_3$ retrievals

The AIRS satellite instrument was the first in a new generation of high spectral resolution infrared sounder instruments flown aboard the National Aeronautics and Space Administration (NASA) Earth Observing System (EOS) Aqua satellite (Aumann et al., 2003, 2020; Chahine et al., 2006; Divakarla et al., 2008). The AIRS radiance data in the 9.6 µm band are used to retrieve column O$_3$ and O$_3$ profiles during both day and night (including the polar night) (Pittman et al., 2009; Fu et al., 2018; Susskind et al., 2003, 2011, 2014). The AIRS V6 Level 3 daily standard physical retrieval products (2003–2018) provide TCO and profiles of retrieved O$_3$. The daily Level 3 products comprise daily averaged measurements on the ascending and descending branches of an orbit with the quality indicators “best” and “good” and are binned into 1° × 1° (latitude × longitude) grid cells. The O$_3$ profile is vertically resolved in 28 levels between 1100 and 0.1 hPa. This makes it possible to compare SCO between AIRS and MLS. Moreover, estimates of the errors associated with cloud and surface properties are part of the AIRS V6 Level 2 standard physical retrieval product, which we used here to discuss further details. Outside of the polar zones (60–90° N and 90–60° S), ascending and descending correspond to daytime (13:30 LST, local solar time) and night-time (01:30 LST) respectively. Hereafter, we refer to “day” and “night” rather than ascending and descending between 60° S and 60° N. In the polar zones, it is inappropriate to use the ascending (descending) mode to define daytime (night-time); therefore, we just compare differences between the ascending and descending mode. AIRS TCO measurements agree well with the global Brewer–Dobson network station measurements with a bias of less than 4% and a root-mean-square error (RMSE) difference of approximately 8% (Divakarla et al., 2008; Nalli et al., 2018; Smith and Barnet, 2019). Analysis of AIRS TCO monthly maps revealed that its retrievals depict seasonal trends and patterns in concurrence with Ozone Monitoring Instrument (OMI) and Solar Backscatter Ultraviolet Radiometer (SBUV/2) observations (Divakarla et al., 2008; Tian et al., 2007).

2.2 MLS stratospheric column of O$_3$ and O$_3$ profile retrievals

The MLS instrument on-board the Aura satellite, which was launched on 15 July 2004 and placed into a near-polar Earth orbit at 705 km with an inclination of 98°, uses the microwave limb-sounding technique to measure vertical profiles of chemical constituents and dynamical tracers between the upper troposphere and the lower mesosphere (Waters et al., 2006). Its orbital ascending mode is at 13:42 LST and the orbital descending mode is at 01:42 LST between 60° S and 60° N. In this study, we use the MLS v4.2x standard O$_3$
product during 2005–2018. Its retrieval uses 240 GHz radiance and provides near-global spatial coverage (82° S–82° N latitude), with each profile spaced 1.5° or ~165 km along the orbit track. This O₃ product includes the O₃ profile on 55 pressure surfaces, and the recommended useful vertical range is from 261 to 0.02 hPa. In addition, it contains an O₃ column, which is the integrated stratospheric column down to the thermal tropopause calculated from MLS-measured temperature (Livesey et al., 2015). Jiang et al. (2007) found that the MLS stratospheric O₃ data between 120 and 3 hPa agreed well with ozonesonde measurements, within 8% for the global daily average. Froidevaux et al. (2008) reported MLS stratospheric O₃ uncertainties of the order of 5%, with values closer to 10% (and occasionally 20%) at the lowest stratospheric altitudes. Livesey et al. (2008) estimated the MLS O₃ accuracy as ~40 ppbv ±5% (~20 ppbv ± 20% at 215 hPa). Expectations and comparisons with other observations show good agreements for the MLS O₃ product, which are generally consistent with the systematic errors quoted above.

3 Results

3.1 AIRS O₃ retrievals’ day–night differences

Figure 1 shows spatial variations in the differences between the AIRS day and night measurements. Generally, over 90% of the globe, AIRS TCO is smaller during night-time than during daytime. The reduction of AIRS TCO over land at night is greater than over oceans depending on the surface type. The seasonal averaged O₃ day-to-night relative differences shown in Fig. 1a–d reveals that AIRS TCO day and night difference variations in Asia, Europe, and North America during winter in the Northern Hemisphere (DJF) are smaller than during summertime (JJA), which is in line with the efficiency of photochemical production between seasons in the Northern Hemisphere. The Sahara Desert shows a maximum difference value during wintertime, when there are large day–night temperature differences. The same phenomenon is observed in Western Australia during summertime. The fact that the presence of a day–night difference appears to correlate with surface infrared emissivity properties of dry desert regions is consistent with Masiello et al. (2014), who discussed the variability of surface infrared emissivity in the Sahara Desert and recommended taking the diurnal variation in the surface emissivity into account in infrared retrieval algorithms.

Figure 1e shows the annual mean large differences of AIRS TCO retrievals over deserts, difference patterns over the oceans associated with the Intertropical Convergence Zone (ITCZ), as well as regions with persistent seasonal subtropical stratocumulus fields. The spatial patterns over land mimic regions with low IR surface emissivity and/or regions where IR surface emissivity exhibits large seasonal variations (Feltz et al., 2018). Figure 1f shows absolute differences between all subsequent pixels in the longitudinal direction. The figure reveals significant non-physical TCO changes (discontinuities) for adjacent land–ocean pixels (visible at coast lines running in the north–south direction). All of these effects are important parameters for the retrieval algorithm, but they bear no physical relation to total O₃. The observed diurnal cycle in AIRS TCO is related to either the measurements or to the algorithm. If the diurnal cycles in AIRS TCO are related to the retrieval algorithm, it has to be caused by the representation of a process in the algorithm having a diurnal cycle; Smith and Barnet, 2019) argue that the issue does not stem from the algorithm but should be taken into account. Hence, the differences shown in Fig. 1 provide strong indications that the largest AIRS day–night TCO differences are dominated by retrieval artefacts. As such, changes are unphysical, and this confirms the hypothesis that clouds and the surface type (land, desert, vegetation, snow, or ice) affect the AIRS TCO retrievals. Note that TCO day–night differences over land could also be (partly) related to clouds.

The AIRS emissivity retrieval uses the NOAA regression emissivity product as a first guess over land. The NOAA approach is based on clear radiances simulated from the European Centre for Medium-Range Weather Forecasts (ECMWF) forecast and a surface emissivity training data set (Goldberg et al., 2003). The training data set used for the AIRS V4 algorithm has a limited number of soil, ice, and snow types and very little emissivity variability in the training ensemble. In the AIRS V5 version, the regression coefficient set has been upgraded using a number of published emissivity spectra (12 spectra for ice and/or snow and 14 for land) blended randomly for land and ice (Zhou et al., 2008). These improvements generated a better emissivity first guess for use with the AIRS V5 and improved retrievals over the desert regions (Divakarla et al., 2008). In AIRS V6, a surface climatology was constructed from the 2008 monthly MODIS MYD11C3 emissivity product and was extended to the AIRS IR frequency hinge points using the baseline-fit approach described by Seemann et al. (2008). Note that AIRS observations with low information content (especially around the poles) will be drawn to the AIRS a priori value. This AIRS a priori value for O₃ is a climatology without diurnal variation. If either the day or night observation has a lower information content than the other, this too can result in a day–night difference. This is probably the reason for the differences in Fig. 1 over pole ice. Nevertheless, using day–night differences for the evaluation of the AIRS V6 O₃ product suggests that further refinements for better surface emissivity retrievals are required and that issues related to cloud cover need to be solved.

3.2 MLS O₃ retrievals’ day–night differences

In order to better understand day–night differences in TCO, we also study day–night changes in the vertical profile of
We find an unexpected polar bias distinguished by the “AscDescMode” flag at high latitudes in Fig. 2c and f. On the one hand, the larger differences between the ascending and descending MLS O₃ profiles at high latitude extend from the stratosphere to the mesosphere; on the other hand, ascending O₃ is smaller than descending O₃ at 10 hPa between 60 and 90° N in Fig. 2c, which is in contrast with the result of other latitudinal bands.

Day and night MLS O₃ profiles distinguished by the “OrbitGeodeticAngle” flag at different latitude bands (30°) between 60° S and 60° N display same results as analysis by “AscDescMode”. Figure 2c and f show that the varieties of ascending and descending MLS O₃ profiles distinguished by
the “OrbitGeodeticAngle” flag at high latitudes are consistent with other regions.

The MLS O$_3$ profile polar bias mentioned above turns out to be related to an inconsistency in the “AscDescMode” flag of the MLS v4.20 standard O$_3$ product between 90 and 60° S and between 60 and 90° N. In version v4.22 and later versions this has been fixed. Figure 3a and c show that there is a clear change in the daily number of ascending and descending pixels on 14 May 2015, which is consistent with the change in MLS SCO in Fig. 3b and d. After 14 May 2015 (using version v4.22), the ascending and descending MLS SCO are much closer. For the MLS O$_3$ profile in Fig. 4, differences between ascending and descending MLS O$_3$ profiles at high latitudes for 2016–2018 are very small. Note that the concept day–night has less physical relevance in polar regions due to the presence of the polar day or night. Outside of polar regions many atmospheric parameters show significant 24 h cyclic changes due to differences in heating and cooling between day and night. Due to Earth’s orbital inclination, 24 h cyclic variations in atmospheric parameters in polar regions are less significant or even absent.

The O$_3$ retrieval algorithm adopted by the MLS v2.2 products has been validated to be highly accurate using multiple correlative measurements, and the data have been widely used (Jiang et al., 2007; Froidevaux et al., 2008). The MLS v3.3 and v3.4 O$_3$ profiles were reported on a finer vertical grid, and the bottom pressure level with scientifically reliable values (MLS O$_3$ accuracy was estimated at $\sim 20$ ppbv + 10% at 261 hPa) increases from 215 to 261 hPa (Livesey et al., 2015). The latest MLS v4.2x O$_3$ profile used in this study, released in February 2015, was generally similar to the previous version. One of the major improvements of MLS v4.2x was the handling of contamination from cloud signals in trace gas retrievals that resulted in a significant reduction in the number of spurious MLS profiles in cloudy regions and a more efficient screening of cloud-contaminated measurements. Furthermore, the MLS O$_3$ products have been improved through additional retrieval phases and a reduction in interferences from other species (Livesey et al., 2015).

### 3.3 Comparison between AIRS and MLS O$_3$ retrievals

Figure 5 presents comparison of yearly and monthly averaged SCO for 2005–2018 observed by AIRS and MLS in three latitude bands. Figure 5 explores the seasonality of either AIRS or MLS SCO day–night differences as well as whether the seasonality in day–night SCO varies in unison over the seasons. Figure 5a shows the 14-year average daytime AIRS SCO (250–1 hPa) and MLS SCO (261–0.02 hPa) between 60° S and 60° N for 2005–2018. The time-averaged MLS SCO column is 260.62 DU and AIRS SCO is 264.24 DU. The average MLS SCO day–night differences for 2005–2018 (0.88 DU) are smaller than the AIRS SCO day–night differences observed for the same time period (5.24 DU). The day–night difference of MLS SCO is 0.79 DU in the mesosphere (10–0.1 hPa) and 0.03 DU in the stratosphere (100–10 hPa). The day–night difference of AIRS SCO is 1.51 DU in the mesosphere (10–1 hPa) and 3.85 DU in the stratosphere (100–10 hPa). Compared with the AIRS SCO day–night differences, the magnitudes of
MLS SCO day–night differences in the stratosphere and in the mesosphere are much smaller. It has been pointed out that errors in temperature profiles and water vapour mixing ratios will adversely affect the AIRS O₃ retrieval. Significant biases (0 %–100 %) may exist in the region between ∼ 300 and ∼ 80 hPa (Wang et al., 2019; Olsen et al., 2017). AIRS O₃ retrievals do not distinguish portions of the O₃ profile as being of different qualities, because all AIRS O₃ channels sense the surface as well as atmospheric O₃. Thus, AIRS O₃ retrievals are compromised if the surface is not well characterized (Olsen et al., 2017). In addition, AIRS SCO retrievals show smaller day–night differences in the polar zones (1–2 DU) than between 60° S and 60° N (4–5 DU). This is related to clouds and the surface type which both affect the AIRS O₃ retrievals as mentioned above. Figure 5b shows the monthly 14-year average daytime AIRS SCO and MLS SCO between 60° S and 60° N for 2005–2018. Seasonal or random changes in clouds and the surface emissivity have a more significant impact on each monthly AIRS SCO retrieval than on the MLS SCO retrieval. Compared with the 60° S–60° N region, surface types in polar zones are less diverse (snow or ice). Therefore, the monthly 14-year average daytime AIRS SCO and MLS SCO in Fig. 5d and f show similar patterns. Figure 5c–f confirm that both MLS and AIRS can catch SCO seasonality at high latitudes. For AIRS SCO in Fig. 5f, the smallest day–night differences occur in September during the Antarctic O₃ hole.

### 3.4 Day–night difference of equatorial Pacific low-O₃ regions

Generally, the Pacific low-O₃ region (TCO < 220 DU), called the zonal wave-one feature (Newchurch et al., 2001; Ziemke et al., 2011), exists all year round. It is caused by lower NOₓ concentrations in this region. Other causes are tropospheric O₃ loss related to higher air temperatures and higher water concentrations. High sea surface temperatures favour strong convective activity in the tropical western Pacific, which can lead to low O₃ mixing ratios in the convective outflow regions in the upper troposphere in spite of the increased lifetime of odd oxygen (Kley et al., 1996; Rex et al., 2014). A further reduction in the tropospheric O₃ burden...
Figure 4. (a) Averaged MLS ozone profile between 261 and 0.02 hPa for 2005–2014 from 60 to 90° N. (b) Averaged MLS ozone profile between 261 and 0.02 hPa for 2016–2018 from 60 to 90° N. Panels (c) is the same as panel (a) but for 90–60° S. Panel (d) is the same as panel (b) but for 90–60° S.

through bromine and iodine emitted from open-ocean marine sources has been postulated by numerical models (Vogt et al., 1999; von Glasow et al., 2002, 2004; Yang et al., 2005) and observations (Read et al., 2008). However, the day–night differences in this region are expected to be small.

Figure 6a and c show that the low-O3 region is mainly located over the western Pacific by AIRS. Rajab et al. (2013) investigated similar low TCO in Malaysia using AIRS data. They found that the highest O3 concentration occurred in April and May, and the lowest O3 concentration occurred during November and December, which is consistent with our results in Fig. 6f. They also found that O3 concentrations exhibited an inverse relationship with rainfall but were positively correlated with temperature. Figure 6b shows that, in addition to the tropical western Pacific, low-O3 regions for MLS appear all over the tropical zone (30° S–30° N) at night. However, Fig. 6d shows that the occurrence frequency and intensity of daytime low-O3 regions by MLS SCO retrievals drastically reduces and exists mainly in tropical western Pacific. In Fig. 6e and f, yearly and monthly averaged AIRS TCO and MLS SCO of the low-O3 regions show no consistency or regularity. The analysis of daytime MLS SCO of the low-O3 regions is based on only a few observations. We cannot distinguish whether it is an algorithm problem or a chemical mechanism that caused this phenomenon. For AIRS, clouds over oceans may have greater impact on the AIRS TCO retrievals at night. For MLS, more active chemical reactions may occur in these low-O3 regions at night.

For past, current, and future monitoring of atmospheric phenomena like the Pacific tropospheric low-O3 area, it is important that observations are sufficiently accurate. The evaluation of day–night differences in both MLS and AIRS has revealed the existence of biases in the satellite data that are large enough in comparison to expected variations and changes in atmospheric O3 that they may hamper the use of these satellite data studying them.

4 Conclusions

Comparison of daytime and night-time AIRS TCO has revealed small but not insignificant biases in AIRS TCO. The differences are likely related to surface type (land, desert, vegetation, snow, or ice) and infrared surface emissivity, especially over regions that exhibit smaller infrared emissiv-
For ocean regions with persistent clouds during day and night (for example, over the ITCZ), Fig. S1 in the Supplement shows that variations in cloud layer height have a greater impact on AIRS TCO day–night differences over land than variations in the cloud fraction.

Our results do not provide much evidence of another possible cause of day–night differences in AIRS TCO: the photochemical diurnal O$_3$ cycle in the lower troposphere and up-
per atmosphere. The strongest diurnal O\(_3\) effects occur in the boundary layer over land due to night-time surface deposition and daytime photochemical O\(_3\) production in the presence of air pollution. In the marine boundary layer, the diurnal O\(_3\) cycle is much weaker due to the absence of air pollution and a general slow O\(_3\) destruction regime (\(\sim 10 \% \text{ d}^{-1}\)). Similarly, in the free troposphere, the diurnal O\(_3\) cycle is also weak due to low O\(_3\) production rates (generally low levels of pollution relevant for O\(_3\) production). Hence, the diurnal O\(_3\) cycle in the free troposphere above 750 hPa is negligible.

**Figure 6.** Spatial and temporal distribution of the low ozone. (a) Location (composite pixel) of the yearly night-time low ozone from 2005 to 2018 for AIRS TCO. Panel (b) is the same as panel (a) but for MLS SCO. (c) Location (composite pixel) of the yearly daytime low ozone from 2005 to 2018 for AIRS TCO. Panel (d) is the same as panel (c) but for MLS SCO. (e) Yearly averaged AIRS TCO and MLS SCO of the low-ozone regions for 2005–2018. (f) Monthly averaged AIRS TCO and MLS SCO of the low-ozone regions for 2005–2018. Uncertainties represent the standard deviation of the measured values.
(Petetin et al., 2016). In summary, any tropospheric photochemical diurnal O$_3$ cycle effect should resemble some correspondence with air pollution. The day–night differences in AIRS TCO clearly do not resemble patterns of surface air pollution (Fig. 1). MLS day–night differences are confined to the mesosphere (1 hPa and higher). As shown in Smith et al. (2014), the lifetime of O$_3$ due to chemistry is strongly altitude dependent (< 20 min in the upper mesosphere above 0.01 hPa). Only in the mesosphere is the chemical lifetime of O$_3$ long enough to see significant differences between average daytime and night-time concentrations. However, the contribution of mesospheric O$_3$ to MLS SCO is negligible. Thus, the mesospheric diurnal O$_3$ cycle will also have a negligible effect on day–night AIRS TCO differences. In addition, Strode et al. (2019) simulated the global diurnal cycle in the tropospheric O$_3$ columns, and their results indicated that the mean peak-to-peak magnitude of the diurnal variability in tropospheric O$_3$ is approximately 1 DU. Figures S2 to S5 also show that the AIRS TCO retrieval artefacts dominate the day–night variability of tropospheric O$_3$ residuals (TOR = AIRS TCO − MLS SCO).

In summary, our analysis has identified evidence and indications that clouds, land surface infrared emissivity, and the sensitivity of satellite measurements to the lower troposphere influence AIRS satellite TCO observations and has pinpointed areas and processes for algorithm improvement.

The MLS v4.2x was very useful for the verification of daytime and nighttime SCO and O$_3$ profiles between 60° S and 60° N. MLS day–night differences in SCO and O$_3$ profiles show that day–night differences are only small (< 1 DU) and are likely to be in the upper stratosphere and mesosphere. However, an inconsistency was found in the “AscDescMode” flag between 60 and 90° N and between 90 and 60° S, resulting in inconsistent profiles in these regions before 14 May 2015. In processor version v4.22 and later versions this issue has been fixed, but as it is a relatively small issue, the MLS data set before 2016 has not been reprocessed (confirmed by Nathaniel J. Livesey, personal communication, 2020).

A case study of day–night differences in O$_3$ over the equatorial Pacific revealed that both AIRS and MLS O$_3$ retrievals have biases in comparison to expected variations and changes. Therefore, our results show that maintaining the quality of the satellite observations of stratospheric O$_3$ is highly relevant.

Data availability. Satellite data sets used in this research can be requested from public sources. AIRS Level 3 data are available online: https://doi.org/10.5067/Aqua/AIRS/DATA303 (AIRS Science Team/Joao Teixeira, 2013a). AIRS Level 2 data are available from https://doi.org/10.5067/Aqua/AIRS/DATA202 (AIRS Science Team/Joao Teixeira, 2013b). MLS Level 2 data can be obtained from https://doi.org/10.5067/Aura/MLS/DATA2017 (Schwartz et al., 2015).

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