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**Ozone inside tropical
deep convective
clouds**

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Ozone mixing ratios inside tropical deep convective clouds from OMI satellite measurements

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We have developed a new technique for estimating ozone mixing ratio inside deep convective clouds. The technique uses the concept of an optical centroid cloud pressure that is indicative of the photon path inside clouds. Radiative transfer calculations based on realistic cloud vertical structure as provided by CloudSat radar data show that because deep convective clouds are optically thin near the top, photons can penetrate significantly inside the cloud. This photon penetration coupled with in-cloud scattering produces optical centroid pressures that are hundreds of hPa inside the cloud. We use the measured column ozone and the optical centroid cloud pressure derived using the effects of rotational-Raman scattering to estimate O_3 mixing ratio in the upper regions of deep convective clouds. The data are obtained from the Ozone Monitoring Instrument (OMI) aboard NASA's Aura satellite. Our results show that low O_3 concentrations in these clouds are a common occurrence throughout much of the tropical Pacific. Ozonesonde measurements in the tropics following convective activity also show very low concentrations of O_3 in the upper troposphere. These low amounts are attributed to vertical injection of ozone poor oceanic boundary layer air during convection into the upper troposphere followed by convective outflow. Over South America and Africa, O_3 mixing ratio inside deep convective clouds often exceeds 50 ppbv which is comparable to mean background (cloud-free) concentrations. These areas contain higher amounts of ozone precursors due to biomass burning and lightning. Assuming that O_3 is well mixed (i.e. constant mixing ratio with height) up to the tropopause, we can estimate the stratospheric column O_3 over clouds. Stratospheric column ozone derived in this manner agrees well with that retrieved independently with the Aura Microwave Limb Sounder (MLS) instrument and thus provides a consistency check of our method.

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1 Introduction

Satellite measurements of emitted thermal infrared (IR) radiances and backscattered sunlight in the ultraviolet (UV) and visible have different sensitivities to cloud properties, particularly for tropical deep convective clouds. Since clouds are highly absorbing/emitting in thermal IR bands, satellite IR sensors are sensitive to the temperature near the tops of these clouds. The IR cloud emissivity approaches unity at a visible optical depth of about 5. Using temperature/pressure profiles, estimates of the cloud top pressure can be obtained with the CO₂ slicing approach (e.g. Menzel et al., 1992).

In contrast, radiative transfer calculations based on optical depth profiles from the CloudSat radar suggest that visible light penetrates much deeper into tropical convective clouds (Vasilkov et al., 2008). Estimates of UV/visible photon path lengths in cloudy conditions can be derived from oxygen (O₂) A-band absorption (e.g. Rozanov et al., 2004), oxygen dimer (O₂-O₂) absorption (e.g. Sneep et al., 2007), or rotational-Raman scattering in the UV (Joiner and Bhartia, 1995). The latter two approaches have been implemented with the Dutch-Finnish Ozone Monitoring Instrument (OMI) flying on NASA's Earth Observing System (EOS) Aura satellite. The retrieved OMI cloud pressures for bright tropical deep convective clouds represent the equivalent Lambertian pressure required to produce the observed amount of oxygen dimer absorption or rotational-Raman scattering. We refer to this cloud pressure as the optical centroid cloud pressure (OCCP).

Vasilkov et al. (2008) showed that OMI OCCPs of deep convective clouds are typically several hundred hPa inside the cloud and are consistent with vertical optical depth profiles derived from the combination of CloudSat radar reflectivity data and MODIS-derived optical depth. Differences between the IR-derived cloud-top pressure and the OCCP can be attributed to enhanced scattering within the clouds, the presence of multiple cloud decks, and sensitivity to cloud geometrical thickness (e.g. Rozanov et al., 2004). In this paper we show that cloud vertical inhomogeneity also contributes to these differences. Specifically, we expand on the work of Vasilkov et al. (2008) by

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examining the vertical dependence of ozone absorption within realistic tropical deep convective clouds.

Our study utilizes the OCCP concept along with measured column ozone from OMI to derive ozone mixing ratios inside tropical deep convective clouds. Two methods of measurement are invoked. The first is similar to the cloud slicing concept proposed by Ziemke et al. (2001). This method assumes a uniform O_3 mixing ratio inside clouds within a confined geographical region. Inside the region an ensemble of pixels is used to derive the in-cloud O_3 mixing ratio from the slope of measured O_3 column versus OCCP. The second method incorporates a residual approach (e.g. Fishman et al., 1990) to determine O_3 in thick clouds by combining O_3 measurements from OMI and the Microwave Limb Sounder (MLS) instrument on Aura. Because ozonesondes are not launched directly into deep convective clouds, these satellite measurements provide a unique means to determine O_3 concentrations inside thick clouds throughout the tropics.

This paper is organized as follows:

- Section 2 details the OMI instrument and data sets.
- Section 3 discusses the sensitivity of UV radiance measurements to O_3 inside deep convective clouds, and
- Section 4 describes the cloud slicing and residual O_3 algorithms.
- Sections 5–7 discuss the retrievals of O_3 in-cloud mixing ratios and their implications.
- Finally, Sect. 8 provides a summary.

2 The Ozone Monitoring Instrument: specifications and data sets

OMI is a UV/VIS nadir-viewing solar backscatter spectrometer that makes daily measurements of Earth radiances and solar irradiances from 270 to 500 nm with a spec-

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tral resolution of approximately 0.5 nm (Levelt et al., 2006). OMI provides near-global coverage with a nadir pixel size of 13 km×24 km in the UV-2 and visible channels. Retrievals of column amounts of O₃ and other trace gases as well as cloud and aerosol properties are provided by observations in these channels.

5 There are two standard OMI total ozone products. Here we use data from the collection 3 OMTO3 version 8.5 (v8.5) product that is based on the Total Ozone Mapping Spectrometer (TOMS) version 8 (v8) total O₃ algorithm. A description of this algorithm may be obtained from the TOMS v8 CD DVD ROM or from the OMI Algorithm Theoretical Basis Document (ATBD) from the TOMS web page http://toms.gsfc.nasa.gov/version8/v8toms_atbd.pdf.
10 Another OMI algorithm based on the Differential Optical Absorption Spectroscopy Method (DOAS) technique gives similar estimates of total column ozone (Kroon et al., 2008b). These total column ozone retrievals have been compared with ground-based data (McPeters et al., 2008) and aircraft-based measurements (Kroon et al., 2008a).

15 There are also two OMI algorithms that determine the OCCP. One is based on O₂-O₂ dimer absorption (Sneep et al., 2008). The other is based on rotational-Raman scattering (RRS) which creates spectral structures in the ratio of backscattered radiance to the solar irradiance known as the Ring effect (Joiner and Bhartia, 1995; Joiner et al., 2004; Joiner and Vasilkov, 2006). The two OMI cloud algorithms provide similar estimates
20 of OCCP for bright clouds although there are small differences due to algorithmic and physical effects (Sneep et al., 2008).

One main difference between the TOMS v8 and the OMTO3 v8.5 algorithms is the treatment of clouds. The TOMS v8 and earlier versions of OMTO3 use a cloud pressure climatology based on thermal infrared cloud-top pressures, while the OMTO3 v8.5 uses
25 OCCP derived with OMI by the RRS method. The use of simultaneously-derived OCCP significantly improves estimates of total column ozone, especially in the presence of bright clouds (Joiner et al., 2006). Here, we use the measured column ozone amounts as opposed to estimates of the total column ozone. The measured column amounts do not include the unmeasured O₃ column below the OCCP (also known as the “ghost

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column”) that is traditionally added to provide estimates of the total column O_3 (from top of the atmosphere to the surface).

3 Sensitivity of UV to O_3 inside deep convective clouds

Most previous radiative transfer simulations of UV/visible light in cloudy conditions have been conducted with vertically homogeneous clouds. Recently, the combination of the CloudSat profiling radar reflectivities (Stephens et al, 2002) with collocated Aqua MODIS optical depth retrievals (Platnick et al., 2003) has been used to produce estimates of cloud optical depth vertical profiles. Vasilkov et al. (2008) used such profiles to validate OMI optical centroid cloud pressures from the rotational-Raman technique in a tropical deep convective complex. This study uses the same optical depth profiles (tropical Pacific, 13 November 2006, CloudSat granule 2906 and OMI orbit 12402) to examine photon penetration and O_3 absorption inside bright vertically-inhomogeneous clouds. We use the Linearized Discrete-Ordinate Radiative Transfer LIDORT-RRS code (Spurr et al., 2007) for all calculations. The details of radiative transfer simulation are described in Vasilkov et al. (2008).

The left side of Fig. 1 shows a single CloudSat/MODIS-derived optical depth profile for a tropical deep convective cloud with a cloud optical thickness of 57. The optical depth per unit length has a large peak near 400 hPa corresponding to the ice portion of the cloud with a smaller peak in the water part of the cloud near 600 hPa.

The right side of Fig. 1 shows O_3 Jacobians computed using

1. the CloudSat/MODIS optical depth profile and
2. a Lambertian-equivalent reflectivity model with the surface placed at the optical centroid cloud pressure.

The Lambertian cloud model produces the same amount of rotational-Raman scattering at 350 nm as that simulated for the CloudSat/MODIS optical depth profile. The

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OMI retrieved optical centroid cloud pressure was very close to the simulated value (Vasilkov et al., 2008). The Jacobians were computed using a 5% perturbation in O₃ optical depth per layer. They are plotted as a change in natural logarithm of the radiance per unit change in layer ozone optical depth normalized to a value of unity in the stratosphere.

The Jacobian indicates enhanced O₃ absorption starting near the physical cloud top and extending well inside the cloud to a pressure near 300 hPa. The sensitivity then decreases near the optical depth peak at ~400 hPa. There is relatively little radiance sensitivity to O₃ absorption in the water portion of this cloud.

Figure 2 similarly shows an optical depth profile with a peak near 600 hPa and a total optical depth of 44.6 along with computed Jacobians. Here, the enhanced ozone sensitivity occurs between ~250–450 hPa with sensitivity dropping off near the optical depth peak at ~600 hPa.

Figure 3 shows that the cloud optical depth profiles in Figs. 1 and 2 are typical of tropical deep convection. In this figure, we computed the mean cloud optical depth profile on 13 November 2006 for two different ranges of cloud optical thickness. The brightest tropical clouds (COT>50) on average have a peak near 400 hPa similar to the profile shown in Fig. 1. Less bright clouds (30<COT<50) peak near the melting point (~600 hPa) similar to the profile in Fig. 2. The cloud-top pressures retrieved with the MODIS CO₂ slicing method for profiles shown in Figs. 1 and 2 (and in general for tropical deep convection) are ~200±25 hPa and close to the physical cloud top as detected by CloudSat. Figure 3 indicates that on average, clouds are relatively thin near the top which allows for the significant photon penetration shown in Figs. 1 and 2.

In order to verify that the Lambertian cloud model produces the same amount of ozone absorption as the CloudSat/MODIS cloud profile (note that it produces the same amount of rotational-Raman scattering at 350 nm by definition), we can compute the change in the radiance resulting from a given value of tropospheric ozone volume mixing ratio (assumed to be constant). This is proportional to the integral over tropospheric layers of the Jacobian for a given layer multiplied by the change in pressure of that layer.

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We find that the Lambertian model predicts the radiances at 323 nm to better than 10%. Therefore, although the Jacobians computed using the CloudSat data and Lambertian models differ somewhat in their vertical structure, the Lambertian model should produce a correct estimate of the ozone volume mixing ratio provided that ozone is well mixed inside the cloud.

4 Retrieval of tropospheric ozone in thick clouds using the ensemble cloud slicing and residual methods

As noted in the Introduction, we invoke two methods for measuring tropospheric O₃ in thick clouds from satellite remote sensing. One is the ensemble cloud slicing method (Ziemke et al., 2001) and the other incorporates the residual technique of Fishman et al. (1990). There are distinct advantages with both of these methods. For cloud slicing, an important advantage is that only one satellite instrument is required, while for the residual method one can obtain very high horizontal resolution measurements.

Ensemble cloud slicing is used to directly estimate ozone volume mixing ratio (VMR) inside convective clouds. Figure 4 shows how this method works schematically. A typical measurement of column O₃ from OMI extends down to the OCCP that lies well below the physical cloud top. Over a selected region (here we use 5° latitude×5° longitude boxes), the slope of measured column O₃ versus OCCP from an ensemble of coincident measurements yields mean VMR. Explicitly, $VMR \text{ (units ppmv)} = 1.27 \times \text{SLOPE}$, where SLOPE is measured in Dobson Units (DU; 1 DU≡2.69×10²⁰ molecules per m²) per hPa. OMI reflectivities $R > 0.6$ coincide generally with deep clouds and scene cloud radiance fractions of 100%. This study uses a more stringent criterion of $R > 0.8$, typical of deep convective clouds, for all calculations.

One feature of the cloud slicing approach is that it is not necessary to have an estimate of the stratospheric column ozone (SCO) amount because it cancels out in the slope of the measured column ozone versus OCCP. However, SCO can be derived from the ensemble cloud-slicing technique by extrapolating the linear fit to the tropopause

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pressure (i.e. it is the intercept of the linear fit). A main assumption for the ensemble cloud slicing method is that both stratospheric column ozone and tropospheric mean mixing ratio do not vary much over the ensemble region. The method in principle may be applied outside the tropics provided that this requirement is adequately satisfied (Ahn et al., 2003; Ziemke et al., 2003).

Tropospheric ozone can also be derived in thick clouds from the residual method in which an estimate of SCO must be provided. The residual method has been applied to TOMS using estimates of SCO from the convective cloud differential (CCD) method (Ziemke et al., 1998). The residual approach has also been applied to OMI using SCO from the Aura Microwave Limb Sounder (MLS) (Ziemke et al., 2006; Schoeberl et al., 2007). The residual method can in principal be applied in both clear and cloudy conditions. In cloudy conditions we can make use of the OCCP concept. In this way an estimate of the tropospheric ozone mean mixing ratio can be obtained for every pixel.

5 Results from the ensemble cloud slicing method

Figure 5 shows scatter plots of measured column O_3 versus OCCP over southern Africa and in the western Pacific for selected days in October 2006. O_3 VMR in the western Pacific is essentially zero while over southern Africa the concentration is ~ 60 ppbv. Enhanced O_3 over southern Africa suggests that O_3 produced from regional pollution including biomass burning, which is largest around September–October each year, reaches the upper regions of the clouds. The elevated O_3 may be caused by other sources, e.g., lightning, and transport effects including the planetary-scale Walker circulation. The near-zero O_3 VMR in the western Pacific is consistent with values measured in the vicinity of convection by ozonesondes (e.g. Kley et al., 1996; Folkins et al., 2002; Solomon et al., 2005). The VMRs for the selected days in Fig. 5 are typical of all other days during October 2006 for both regions.

Zonal variability of in-cloud VMR is shown in Fig. 6 for October 2006. Data are for $5^\circ \times 5^\circ$ regions averaged over the month from daily measurements and for OMI include

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a 3-point running average smoothing along longitude. The two latitude bands (10° N–15° N and 5° N–10° N) shown in this figure lie within or near the Inter-tropical Convergence Zone (ITCZ) where deep clouds occur frequently. Figure 6 also shows the corresponding zonal variability of mean equivalent VMR for nearly cloud-free scenes (reflectivity $R < 0.2$). This cloud-free VMR is derived from tropospheric column O_3 (measured in DU) using the CCD method. (The CCD method and its validity are discussed in Sect. 6.) The conversion from column O_3 to mean equivalent VMR (in units ppmv) is obtained by dividing the tropospheric O_3 column in DU by total differential pressure (surface pressure minus the tropopause pressure) in hPa followed by multiplying this number by 1.27 (e.g. Ziemke et al., 2001, and references therein).

Figure 6 indicates that the zonal variability in O_3 VMR inside the deep clouds is qualitatively similar to cloud-free scenes. However, absolute values of O_3 VMR in the cloud-free scenes are larger than the corresponding values inside thick clouds most everywhere. This is particularly apparent in the Pacific region where the O_3 VMR values under cloudy conditions are close to zero.

There are two components of information obtained from the linear fit of measured column O_3 versus OCCP with the ensemble cloud slicing method. Mean VMR is derived from the slope, and SCO is estimated from the intercept. If we assume that O_3 between OCCP and the tropopause is well mixed (i.e. constant VMR), we can then estimate SCO by extrapolating the linear fits to the tropopause pressure (taken everywhere to be 100 hPa for tropical latitudes). The column amount at 100 hPa is then an estimate of SCO.

Figure 7 compares zonal variability of SCO for October 2005 and 2006 over thick clouds derived by cloud slicing using OMI (asterisks) with that derived from MLS under all conditions (thick curves). The Aura MLS measurements of SCO are described by Ziemke et al. (2006). The values shown for both OMI and MLS were first calculated from daily measurements in $5^\circ \times 5^\circ$ grid boxes and then averaged over the month. The OMI data include a 3-point running average smoothing along longitude.

SCO from MLS and OMI in Fig. 7 agrees remarkably well with mean differences of

~1 DU and zonal RMS differences of ~2–3 DU. Such good agreement can only be obtained if the measured ozone and the OCCP concept are correct. If we had asserted that OMI does not measure O₃ below the physical cloud-tops as assumed previously by Ziemke et al. (1998), then we would still get the correct SCO in the Pacific because

5 O₃ inside these thick clouds is negligibly small. However, SCO values derived elsewhere by cloud-slicing would be larger and show longitudinal variability that would be inconsistent with other satellite and sonde data. Hence, this SCO comparison provides an independent validation of the in-cloud VMR amounts (and thus of both the OCCP and O₃ column from OMI) derived in this study.

10 6 Results from the OMI/MLS residual method

Figure 8 shows the measured column O₃ from OMI (top panel) and SCO from the MLS instrument (bottom panel) for October 2005. Both column measurements were obtained from daily level-2 footprint measurements binned to 1° latitude×1.25° longitude resolution. Here we use OMI measured column O₃ values for pixels containing

15 bright clouds with scene reflectivity $R > 0.8$ and OCCP <750 hPa with a 3-point running average smoothing along longitude and latitude. As in Ziemke et al. (2006), MLS SCO was determined from spatially interpolated daily data averaged over each month. The validation of MLS O₃ profile data is discussed by Froidevaux et al. (2008).

The two panels in Fig. 8 indicate nearly equal values of column O₃ (~220 DU) from

20 OMI and MLS in the Pacific. However, the column amounts differ significantly outside of this region. The differences are due to a tropospheric O₃ contribution lying between tropopause and OCCP which varies from nearly zero DU in the Pacific to about 30 DU over southern Africa and South America. SCO over the entire tropical region (lower panel) is nearly constant with values ~225–230 DU.

Figure 8 provides a foundation for the convective cloud differential (CCD) method that was originally based on two assumptions:

1. UV measuring instruments such as TOMS measure the ozone above the tops of

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highly reflective clouds; therefore SCO can be estimated using pixels that contain highly reflective clouds (reflectivity $R > 0.8$) with cloud tops near the tropopause level in the tropical Pacific region, and

2. SCO is to first order zonally invariant within the latitude range 15° S to 15° N.

5 An important property of the CCD technique, which follows from the first assumption, is that cloud height information is not needed to derive tropospheric column ozone. The second assumption was based upon many years of O_3 measurements from satellites and ozonesondes (e.g. Ziemke et al., 1998; Thompson et al., 2003). Tropospheric column ozone at any latitude and longitude is then calculated by differencing low reflectivity ($R < 0.2$) gridded total column ozone and high reflectivity ($R > 0.8$) SCO from the Pacific region. Although we have shown here that the first assumption is not correct, the CCD method still provides an accurate estimate of SCO because the O_3 VMR
10 in the tropical Pacific inside deep convective clouds is exceedingly small to essentially zero. This study provides a new interpretation of the CCD method and shows why it worked as implemented previously.

15 The measured column O_3 from OMI in Fig. 8 (top) is identically the sum of SCO and tropospheric column O_3 lying between the tropopause and the OCCP. Thus taking the difference of the two maps (top panel minus bottom panel in Fig. 8) yields column O_3 lying between OCCP and the tropopause. This column amount can then be converted to mean VMR as described earlier for Fig. 6. As generated in this fashion, maps of upper tropospheric O_3 are shown in Fig. 9 for July and October 2006. OCCP values were constrained to lie between 250 hPa and 550 hPa to estimate mean O_3 concentrations in the upper troposphere. As in Fig. 8 column amounts were binned to 1° latitude \times 1.25° longitude resolution from daily level-2 footprint measurements and include a 3-point
20 running average smoothing along longitude and latitude.

25 For July and October 2006 there are some pixels in the eastern and western Pacific that have negative VMRs of a few ppbv. As discussed by Ziemke et al. (2006), SCO from MLS is adjusted to OMI CCD measurements of SCO. The small localized

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values of negative O_3 in the Pacific could be explained by yet unresolved measurement/calibration differences between OMI and MLS for bright cloud scenes.

7 Discussion

As mentioned in Sect. 5, studies of tropospheric O_3 from ozonesonde measurements have shown occurrences of low to near-zero concentrations of O_3 in the tropical upper troposphere (Kley et al., 1996; Folkins et al., 2002; Solomon et al., 2005). These low values are largely attributed to convective lifting of low amounts of O_3 from the marine boundary layer into the upper troposphere. In pollution-free oceanic regions it is not uncommon for O_3 in the marine boundary layer to be only a few ppbv owing to reactions involving hydrogen radicals OH and HO_2 (e.g. Solomon et al., 2005, and references therein). Some studies have also suggested the possibility of in-cloud photochemical destruction mechanisms of O_3 (e.g. Zhu et al., 2001; Barth et al., 2002; Liu et al., 2006). The previous inferences of near-zero concentrations of O_3 in the tropical upper troposphere have been based almost entirely on ozonesondes and generally limited to a few isolated sites in the vicinity of (but not inside) deep convective cloud systems.

It is conceivable that the O_3 mixing ratio distributions in Fig. 9 are largely a manifestation of air in the lower troposphere (including the planetary boundary layer) that is injected upward into the upper portions of clouds during convection. Exceedingly low O_3 lies over the western Pacific extending across the dateline into the eastern Pacific along the ITCZ. Low amounts are also present in the Indian Ocean.

In contrast to the Pacific, elevated O_3 occurs over the large land masses of Africa and South America in cloud free conditions as well as in the presence of deep convective clouds. The O_3 concentrations exceed 50 ppbv over much of northern Africa in these months, and also southern Africa in October. These enhanced concentrations suggest a much more polluted environment including the boundary layer. There is also substantial amounts of tropospheric O_3 for thick clouds extending from west of Central America eastward across the Atlantic to northern Africa in all months. The O_3 VMRs

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for these thick clouds over ocean have values of $\sim 30\text{--}40$ ppbv.

October 2006 coincided with a strong El Niño event which contributed to the elevated tropospheric O_3 seen in the western Pacific (Logan et al., 2008). While the dynamical changes (i.e. reduced convection during El Niño) in the western Pacific increased O_3 concentrations, some of the increases were also caused by lightning and biomass burning in Indonesia in October.

8 Summary

In this paper we have developed a new technique for deriving tropospheric ozone mixing ratios inside deep convective clouds and the stratospheric column ozone (SCO) lying above these clouds. The technique makes use of the concept of the optical centroid cloud pressure (OCCP) obtained from the effects of rotational-Raman scattering in the OMI UV spectra. We applied a radiative transfer model (LIDORT) to study UV photon penetration and O_3 absorption within deep convective clouds. The calculations are based upon realistic cloud optical depth profiles in tropical deep convective clouds from CloudSat and MODIS data. The calculations show that tropospheric O_3 sensitivity at UV wavelengths for deep convective clouds is largest within the upper portion of these clouds with significant sensitivity above the cloud optical depth peak, but still well below the physical cloud top. Cloud optical depth typically peaks near 400 hPa for the brightest clouds and near 600 hPa for less bright convective clouds.

Assuming well-mixed tropospheric O_3 (i.e. constant mixing ratio), we estimated SCO above thick clouds. The close agreement with SCO derived independently from Aura MLS and OMI provides an internal consistency check of measured O_3 column amounts, the derived optical centroid cloud pressures, and thus the derived in-cloud O_3 mixing ratios. Our study indicates very low and even near-zero O_3 in the middle-to-upper troposphere over much of the tropical Pacific region. These values are consistent with previous studies based on ozonesonde measurements made in close proximity to regions of convective activity. In contrast to the Pacific region, O_3 mixing ratios are

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significantly elevated (often exceeding 50 ppbv) over the large landmasses of Africa and South America. The elevated concentrations are comparable to O₃ mixing ratios in clear sky conditions and do not appear to be much perturbed by the presence of deep convection.

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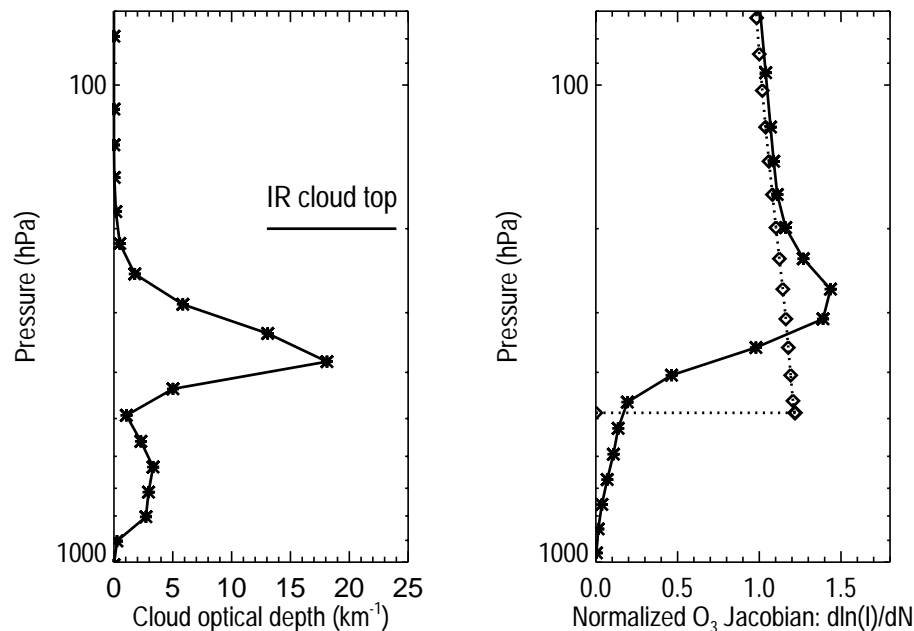


Fig. 1. Left: Cloud optical depth profile from CloudSat/MODIS and approximate MODIS cloud-top pressure from a tropical sounding on 13 November 2006. Right: Normalized ozone Jacobians (partial derivative of the natural logarithm of the radiance at 322.9 nm with respect to the ozone optical depth – normalized to unity in the stratosphere) computed using CloudSat/MODIS optical depth profile (solid line with stars) and for a Lambertian cloud located at the optical centroid pressure (dotted line with diamonds).

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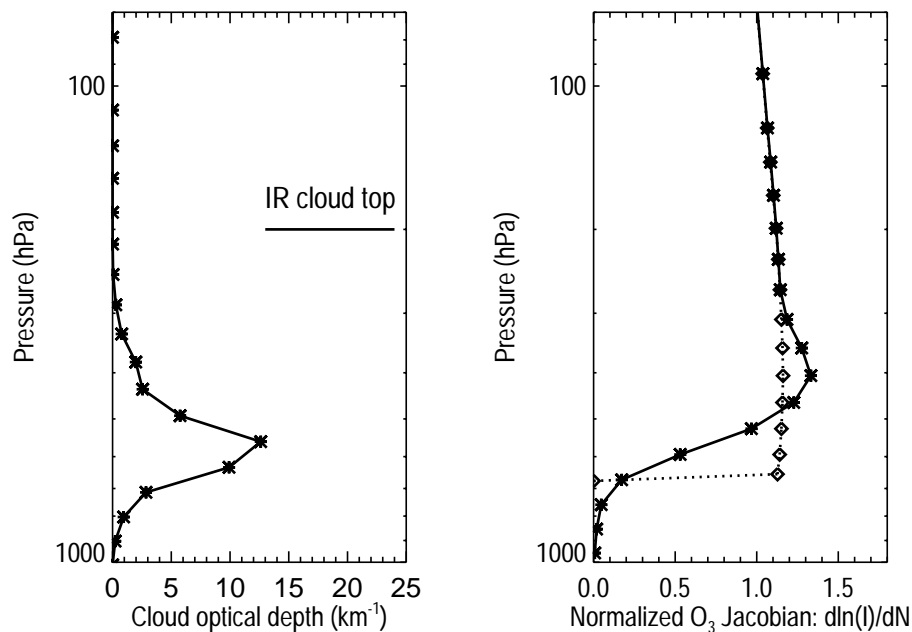


Fig. 2. Similar to Fig. 1, but for a different CloudSat/MODIS pixel.

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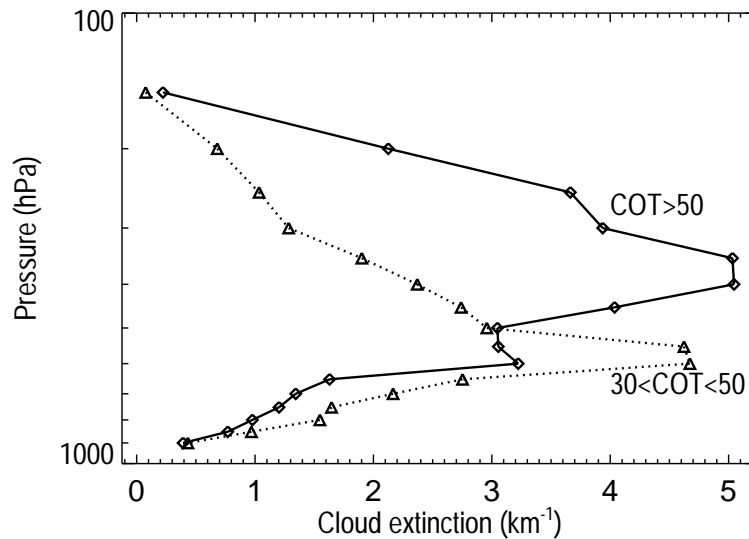


Fig. 3. Mean tropical cloud optical depth profiles derived from CloudSat/MODIS data on 13 November 2006 for different ranges of cloud optical thickness (COT).

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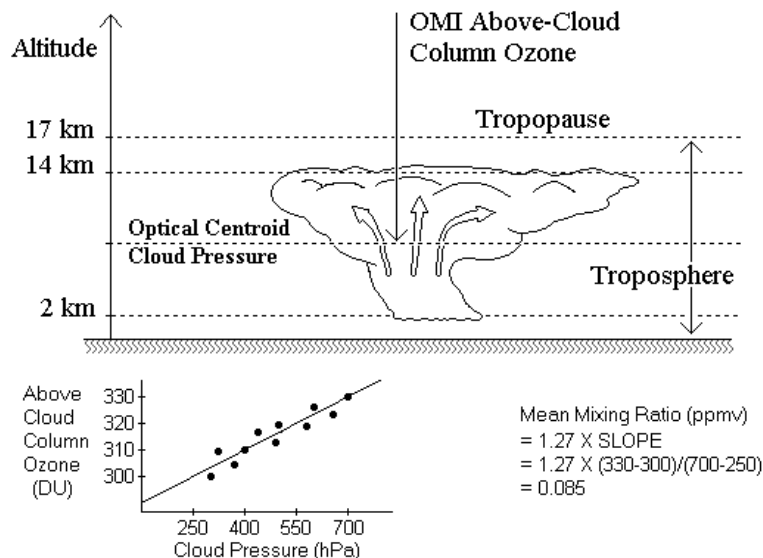


Fig. 4. Schematic diagram illustrating the ensemble cloud-slicing method. The figure shows that a satellite UV instrument is sensitive to the O_3 column from the top of the atmosphere down to the OCCP altitude which may lie several hundred hPa below geometrical cloud top. The lower half of the figure illustrates that using an ensemble of such measurements over a fixed region, mean volume mixing ratio can be determined from the slope of column O_3 plotted versus OCCP.

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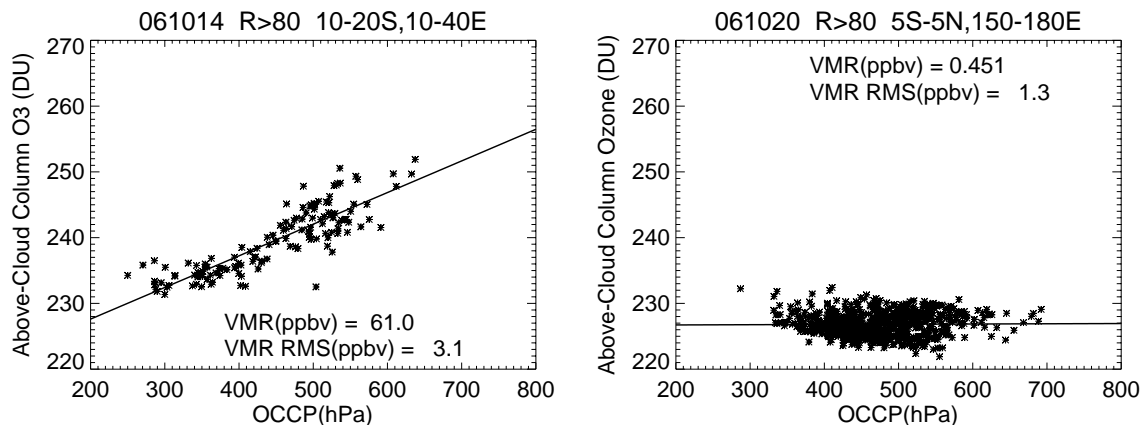


Fig. 5. Daily scatter plots of O₃ column versus OCCP in October 2006. Left panel: southern Africa on 14 October 2006. Right panel: western Pacific on 20 October 2006. Both mean VMR and its RMS uncertainty (2σ , in ppbv) from the regression line fitting are shown.

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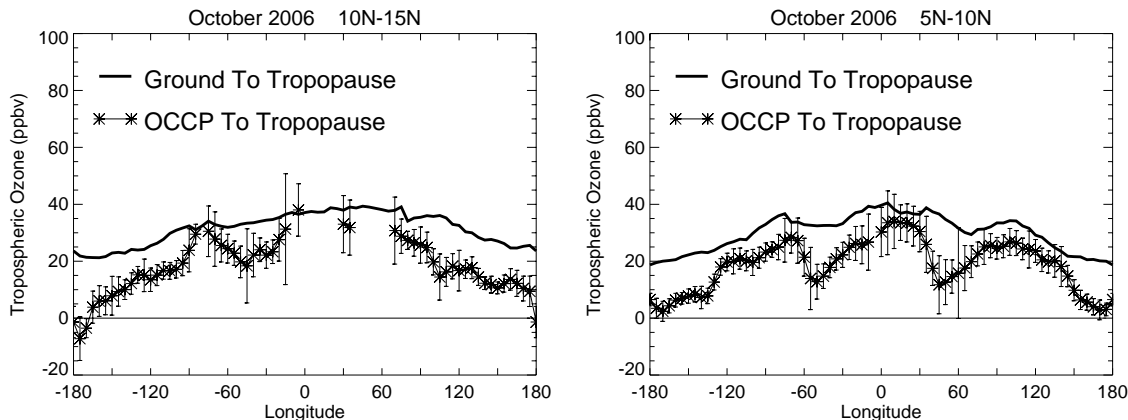


Fig. 6. Tropospheric O_3 VMR from OMI as a function of longitude for thick-clouds (indicated by asterisks) in two tropical latitude bands in October 2006. Values for VMR include $\pm 2\sigma$ regression uncertainties (values for σ represent standard error of the mean). The solid curves represent VMR for nearly cloud-free sky (reflectivity < 0.2) from the CCD method (see text). Data are for $5^\circ \times 5^\circ$ regions averaged over the month from daily measurements and for OMI include a 3-point running average smoothing along longitude.

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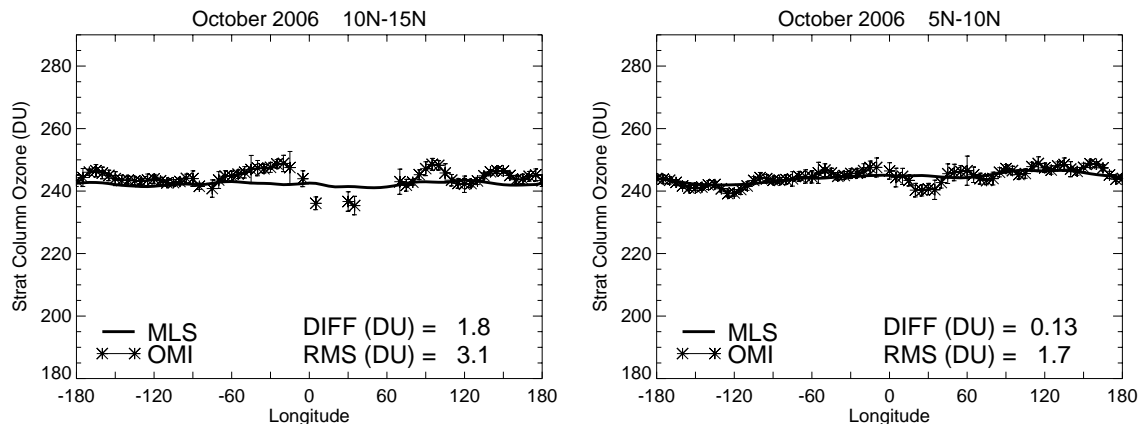


Fig. 7. SCO (in DU) versus longitude in two tropical latitude bands for October 2006. Asterisks: SCO derived from OMI using thick clouds; Solid lines: SCO from MLS. Error bars are $\pm 2\sigma$ uncertainties (values for σ represent standard error of the mean). Mean difference (OMI minus MLS) and zonal RMS of the difference in DU are listed in both panels. Data are for $5^\circ \times 5^\circ$ regions averaged over the month from daily measurements and for OMI include a 3-point running average smoothing along longitude.

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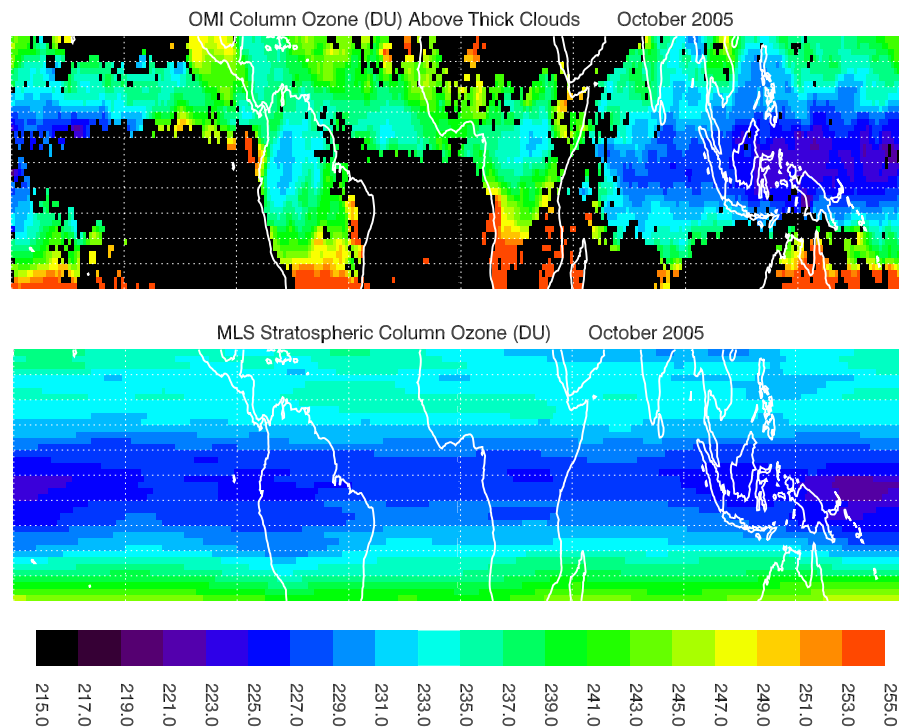


Fig. 8. Measured column O_3 from Aura OMI (top panel) and SCO from the Aura MLS instrument (bottom panel) averaged for October 2005. Column amounts for both OMI and MLS were binned to 1° latitude $\times 1.25^\circ$ longitude resolution from daily level-2 footprint measurements. OMI O_3 was filtered for conditions of bright clouds (scene reflectivity $R > 0.8$), $ECP < 750$ hPa, and includes a 3-point running average smoothing along longitude and latitude. SCO was determined from spatially interpolated daily MLS data which were then averaged monthly. Black pixels over ocean and northern Africa in the top panel indicate no bright clouds in these regions for the entire month.

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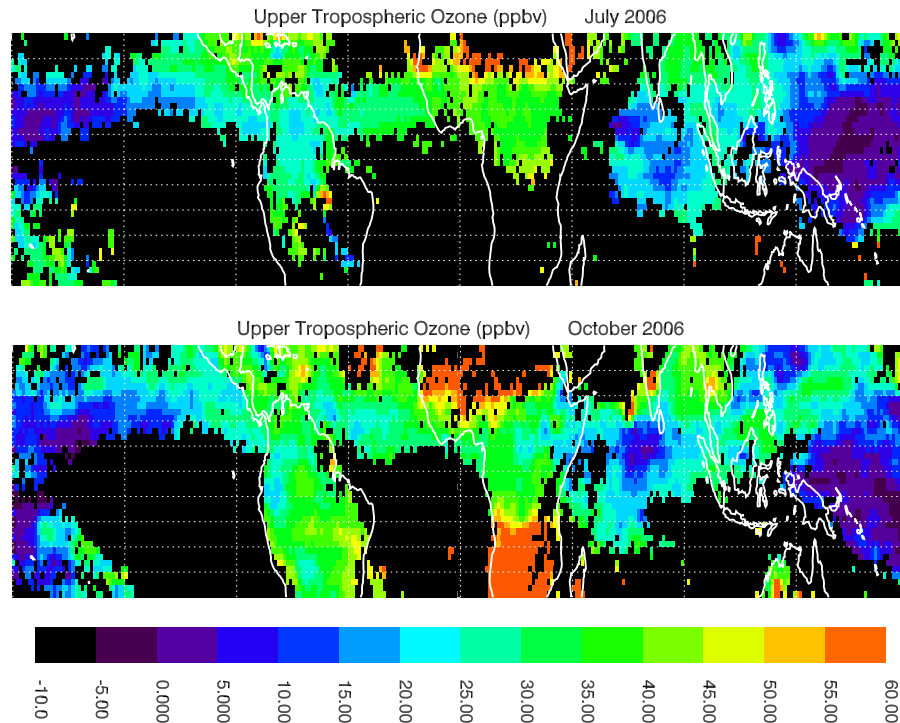


Fig. 9. Estimated mean upper tropospheric O_3 VMR (ppbv) for July 2006 (top) and October 2006 (bottom). OCCP values were constrained to lie between 250 hPa and 550 hPa. Column amounts were binned to 1° latitude \times 1.25° longitude resolution from daily level-2 footprint measurements which include a 3-point running average smoothing along longitude and latitude. Black pixels over ocean and northern Africa indicate no bright clouds in these regions for the entire months.

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