Atmos. Chem. Phys. Discuss., 10, 1939–1956, 2010
www.atmos-chem-phys-discuss.net/10/1939/2010/

Atmospheric Chemistry and Physics Discussions

This discussion paper is/has been under review for the journal Atmospheric Chemistry and Physics (ACP). Please refer to the corresponding final paper in ACP if available.

Aerosol-induced changes of convective cloud anvils produce strong climate warming

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Received: 7 January 2010 – Accepted: 12 January 2010 – Published: 25 January 2010

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Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

The effect of aerosol on clouds poses one of the largest uncertainties in estimating the anthropogenic contribution to climate change. In contrast, even small human-induced perturbations in cloud coverage, lifetime, height or optical properties can change the instantaneous radiative energy flux by hundreds of watts per unit area, and this forcing can be either warming or cooling. Clouds and aerosols form a complex coupled system that, unlike greenhouse gases, have relatively short lifetime (hours to days) and inhomogeneous distribution. This and the inherent complexity of cloud microphysics and dynamics, and the strong coupling with meteorology explain why the estimation of the overall effect of aerosol on climate is so challenging.

Here we focus on the effect of aerosol on cloud top properties of deep convective clouds over the tropical Atlantic. The tops of these vertically developed clouds consist of mostly ice and can reach high levels of the atmosphere, overshooting the lower stratosphere and reaching altitudes greater than 16 km. We show a link between aerosol, clouds and the free atmosphere wind profile that can change the magnitude and sign of the overall climate radiative forcing.

This study demonstrates the deep link between cloud shape and aerosol loading and that the overall aerosol effect in regions of deep convective clouds might be warming. Moreover we show how averaging the cloud height and optical properties over large regions may lead to a false cooling estimation.

1 Introduction

Increase in aerosol concentration may invigorate convective clouds through the particles' role as cloud condensation nuclei (CCN) and the related feedbacks ignited by changes of the clouds' initial droplet size distribution (Andreae et al., 2004; Koren et al., 2005). The proposed chain of events links initial changes in the size distribution of the cloud's droplets to changes in the net updraft velocities, droplet growth rates, cloud

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vertical development and precipitation in the warm, mixed and cold phases.

Previous studies show clear correlations between cloud coverage, lifetime (Albrecht, 1989; Jacobson et al., 2007; Kaufman and Koren, 2006) and optical properties (Twomey, 1977; Roesenfeld, 2000). It was shown from satellite data analysis and in situ measurements, that for convective clouds, increase in aerosol loading is associated with taller invigorated clouds, larger cloud fraction and more extensive ice portions. These associations are found over the tropical Atlantic (Koren et al., 2005; Jenkins and Pratt, 2008; Jenkins et al., 2008), Europe (Devasthale et al., 2005), North and South America (Andreae et al., 2004; Koren et al., 2008a; Lindsey and Fromm, 2008), and appear for all types of aerosol particles: biomass burning smoke, urban/industrial aerosol and desert dust. Numerical modeling studies show aerosols penetrating cloud base of maritime clouds dramatically increasing the amount of supercooled water, as well as the ice contents and vertical velocities (Khain et al., 2008a,b). Aerosol particles can also affect convective cloud properties when ice nucleating particles are entrained directly into the ice portion of the cloud (Fridlind et al., 2004).

Deep convective clouds consist of the cloud tower (or towers) where most of the convection occurs and the cloud anvil (or anvils), which is a thinner layer of mostly ice particles extending away from the cloud towers. Anvils form when part of the cloud spreads out below a stable atmospheric (inversion) layer, such as the tropopause. The anvil horizontal area can be much larger than the tower area, as the inversion layer above it acts as a ceiling, blocking the anvil cloud elements from expanding upwards and forcing horizontal expansion. The anvil elements are further spread by the horizontal winds below the inversion level (Fig. 1a).

Over the tropical Atlantic, convective cloud liquid water optical depth and ice optical depth respond differently to increase in aerosol loading (measured by the aerosol optical depth, AOD) as observed from satellites (Koren et al., 2005). As expected, due to changes in the droplets size-distribution and warm rain suppression, liquid cloud optical depth increases with AOD, however, ice cloud optical depth decreases. Here by analyzing horizontal wind profiles from radiosonde data from several stations in the

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tropics and subtropics (NOAA/ESRL Radiosonde Database), cloud and aerosol level-3 MODIS-daily data (Platnick et al., 2003; Remer et al., 2008), we show that the clear reduction in the average ice optical depth with increase in aerosol loading is due to a strong increase in the ice anvil area. We then estimate the anvil enhancement consequences to the climate radiative balance (Ricchiazzi et al., 1998). The study area focuses on the tropical Atlantic (0° N to 14° N; 18° W to 45° W) of June-July-August 2007.

2 Analysis and results

We introduce the tower to anvil area ratio (TAR), and show that while clouds are invigorated by increasing aerosol loading and reach higher levels in the atmosphere, this ratio decreases due to the associated increase and spread of the anvil area. The anvil area increases both due to the enhanced convection and stronger horizontal winds in the upper troposphere. Figure 1b shows the average horizontal wind speed of the free atmosphere (above the boundary layer) as a function of pressure level. A consistent sharp increase with height occurs from 300 to 150 hPs (∼9 to 15 km) and then decreases towards the base of the tropopause (100 hPa). This behavior is consistent for the wide variety of stations analyzed (day and night, land and ocean, inside and outside the tropics, for all seasons). The high level jet contributes to spreading and diluting the anvils, expanding the total cloud cover and decreasing the TAR. Indeed, Fig. 1c clearly shows that the average cloud fraction (each point is an average of 60 one-degree MODIS pixels) increases as a function of the cloud top height. The change in slope at ∼850 hPa marks the transition from the lower (marine boundary layer) to the free-upper troposphere where the clouds have larger average fraction. At 300 hPa almost complete overcast conditions are reached, corresponding to the bottom level of the upper level jet.

How does aerosol loading affect the dependence of cloud fraction and height? The cloud fraction data was sorted according to the aerosol optical depth (AOD) and divided into 3 equally-sampled groups with mean AOD of each group equal to 0.13, 0.24, 0.41, respectively. Then cloud fraction was plotted as a function of the pressure level for each group (Koren et al., 2008a). While the cloud fraction of all 3 groups increases as a function of the pressure level, a clear shift of the polluted clouds towards larger fraction is shown for each pressure level. Moreover the invigoration effect is also shown as the polluted clouds are taller (Fig. 1d).

MODIS retrieves detailed aerosol and cloud properties with resolution of 1 to 10 km. The data are averaged into a daily 1° resolution grid allowing for simultaneous observations of aerosols in cloud-free regions and clouds in the cloudy regions of the grid box when not completely overcast. To save the information of the higher resolution retrievals (1 km^2 for clouds and 10 km^2 for aerosol) histograms of the original cloud and aerosol properties are provided (King et al., 2003). The histograms of the ice cloud optical depth (τ) were analyzed as a function of the AOD over the study area. After inspection of many high resolution anvil optical depth data over the region, we divided the histograms into two regimes: anvil (τ<10) and tower (τ>10). The Tower to Anvil area ratio (TAR) is the averaged number of pixels in the tower regime (N_t) divided by the number of pixels in the anvil regime (N_a):

$$\mathrm{TAR} = \frac{N_\mathrm{t}}{N_\mathrm{a}} \quad (1)$$

The average cloud optical depth τ_c is weighted by the horizontal coverage area of the anvil optical depth (τ_a) and the tower optical depth (τ_t):

$$\tau_\mathrm{c} = \frac{N_\mathrm{t}\tau_\mathrm{t} + N_\mathrm{a}\tau_\mathrm{a}}{N_\mathrm{t} + N_\mathrm{a}} = \frac{\mathrm{TAR}\tau_\mathrm{t} + \tau_\mathrm{a}}{\mathrm{TAR} + 1} \quad (2)$$

As the TAR becomes smaller, τ_a contributes more to the total cloud optical depth τ_c. Figure 2a shows that the TAR decreases by more than half (from 0.59 to 0.26) for an average AOD variation from clean 0.11 to polluted 0.49. Furthermore, Fig. 2b shows that the average ice cloud fraction increases significantly as a function of the AOD and that most of the increase is in the anvil area. Figure 2c shows that as the aerosol loading increases, the anvil height increases and the anvil optical depth decreases significantly.

The proposed chain of events suggested here is the following: aerosols serving as CCNs invigorate convection and increase the cloud vertical extent into the upper troposphere (Figs. 1d and 2c). The taller clouds now reach higher in the atmosphere and spread horizontally, aided by the stronger winds aloft (Fig. 1b), and result in higher (mostly anvil) cloud fraction (Figs. 1d and 2b). As the wind shear stretches the anvils to cover a larger area, the ice path is diluted in the process, resulting in a smaller TAR (Fig. 2a) and a decrease in anvil optical depth (Fig. 2c).

How does such a chain of events affect the climate radiative forcing? Clouds cool the atmosphere by reflecting back to space part of the incoming shortwave solar radiation. They warm the atmosphere by absorbing longwave radiation emitted from the surface and lower atmosphere, and therefore, reduce the thermal energy loss to space. These two different radiative processes depend on different cloud properties. The longwave cloud radiative effect depends on cloud temperature and emissivity. Cloud temperature is linked to the height and physical thickness of the cloud, and the local profile of air temperature. Cloud emissivity is linked to cloud liquid water content, but saturates quickly so that cloud emissivity is mostly constant for a wide range of cloud optical depths. The important point is that higher clouds with colder cloud top temperatures will emit less longwave radiation to space, thereby warming the atmosphere. While on the shortwave side, cloud height plays a minor role and the liquid water/ice content, cloud optical depth, droplet size distribution and thermodynamic phase are the critical cloud parameters that determine cloud reflectance. Optically thicker clouds with smaller droplets/ice particles will reflect more shortwave radiation back to space. The net radiation effect is a superposition of the longwave and shortwave processes.

Figure 2d introduces the τ-Z space of a deep convective tropical cloud. It shows the daily average net radiative forcing calculated using the SBDART radiation transfer model (Ricchiazzi et al., 1998) for a typical tropical temperature profile, as a function of the cloud optical depth and height. Cloud shortwave reflectance depends not only on the cloud optical depth (as presented in the τ-Z space) but also on the clouds' droplet effective radius and phase (water or ice). For a first approximation, we choose a combination of parameters that will accentuate the cloud cooling regime. The transition to ice is delayed until close to $-40\,^{\circ}$C for complete freezing and the effective radius profile is skewed towards large droplets and crystals, from 20 µm in the lower cloud to 40 µm at higher levels. Other calculations done for different droplet sizes and water/ice clouds show similar patterns but with warmer total forcing (see Appendix A).

Such forcing representation reveals the opposing effects of the solar vs. the thermal spectral regimes. Moving left in the diagram – towards lower τ, or up – towards higher altitudes indicates transition to warming. The τ-Z space also demonstrates that the anvil and the tower have to be treated independently for the forcing calculations. Averaging τ (tower+anvil) will often result in τ values >10, corresponding to cooling for clouds below 10.5 km. In reality most of the sky is covered by anvils (Fig. 2b) with $\tau \sim 4$ and cloud-tops higher than 10.5 km creating an overall warming effect (Fig. 2c). Thus, as additional aerosol causes invigoration and enhanced anvil development, we see a propensity towards a radiative warming. Traditionally, aerosol indirect effects are thought to be primarily cooling. Here we show that by enhancing anvil development, the result can be significant warming instead.

3 Discussion

When studying cloud aerosol interactions from observations two inherent problems will always surface namely 1) cloud contamination effects on the aerosol retrievals and 2) cause and effect – separation of true net aerosol effects on clouds from meteorological effects driving both aerosol and clouds.

On one hand one wants to measure aerosol as close as possible to clouds in order to truly reflect the relevant aerosol (that interacts with the clouds) properties. On the other hand, one asks for very accurate measurements of aerosol loading and properties

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which is very difficult to achieve in the vicinity of clouds (Koren et al., 2007, 2008b; Charlson et al., 2007; Wen et al., 2007). Even if assuming that one can measure correctly clouds and aerosol in a cloud field, the next challenge is due to the strong coupling of both to meteorology (environmental properties). Are the shown trends between aerosol loading and clouds anvils due to real aerosol effect on clouds or does the meteorology drive the changes in both aerosol and clouds properties?

The question of decoupling the meteorology from the net aerosol effect was tackled in Koren et al. (2005). There they suggested that vertical velocity form reanalysis model can serve as a good proxy for deep convection. By dividing the data to subsets with similar vertical velocity (magnitude and sign) and analysing each set separately they showed that the aerosol effect of invigoration of the convective clouds is true for any of the subsets.

A recent study was conducted with the objective of providing deeper and detailed answers to these two key questions over the same study area (Koren et al., 2010). It shows that even after rejection of most of the pixels that may contain cloud contamination, the trends between aerosol and clouds are kept. Moreover, cloud properties as a function of aerosol loading show similar trends when using GOCART aerosol transport model (Chin et al., 2000) as a measure for the aerosol loading (instead of satellites measurements). In the study, 280 meteorological variables were analysed to see which variable represents best the properties of deep convective clouds measured from space. The selected variables helped to divide the data into subsets that represent different meteorological conditions. Convective clouds and aerosol analysis were conducted per each meteorological state showing similar aerosol effects for all the sets.

Additional limitation of this study is related to the necessity of having both clouds and aerosol information over the same region. As cloud cover approaches full overcast conditions, aerosol retrievals become rare. Such is the case with the most vigorous convection having the broadest spread of anvils. This may suggest that the present analysis underestimates the number of clouds invigorated by aerosols into the higher atmosphere warming regime of Fig. 2d, making a quantitative estimate of the total global effect extremely difficult (see Appendix B).

In numerous climate studies cloud aerosol interaction is shown to have the potential to be a key contributor to the radiative forcing (see Heintzenberg and Charlson, 2009 for detailed information). Small changes in the cloud shape, structure or lifetime can change significantly the local radiative balance.

Here we suggest that since the wind speed increases significantly in the upper atmosphere, if indeed aerosols invigorate deep convective clouds, such invigoration will have the following consequences: invigorated convection associated with increased aerosol loading results in expanded anvil spatial coverage and reduced anvil optical depth in higher levels of the atmosphere. These thinner, colder clouds might induce strong climate warming. Such a pathway of aerosol effects on clouds, leading to a positive climate forcing, rather than the accepted over all cooling, holds serious consequences to estimates of future climate change.

Appendix A

More on the τ-Z space:

Clouds interact with the electromagnetic radiation in both the solar and thermal range. In the thermal range, the cloud radiative effect depends on two parameters: cloud temperature and emissivity. Cloud temperature is linked to the height of the cloud and the local profile of air temperature. Cloud emissivity is linked to cloud liquid water content, and saturates as liquid water increases. Cloud reflectance, in the solar range, depends mostly on the water amount, the droplets size-distribution and the thermodynamic phase. Liquid water content is not measured directly from remote sensing but can be estimated as the product of the cloud optical thickness τ and cloud effective radius r_e, (which is the ratio of the 3rd and 2nd moments of the droplet size distribution). Both τ and r_e are retrieved simultaneously from spectral reflectance measurements using one visible channel and one channel in the mid-infrared (Nakajima and King, 1990; Platnick et al., 2003).

Cloud r_e will be smaller with higher concentrations of cloud condensation nuclei (CCN), but will increase with height in the cloud. Cloud freezing (thermodynamic phase) depends on the temperature, the ice nucleus concentration and the CCN concentration. It was shown that higher CCN concentrations delay the freezing levels until colder temperatures are reached higher in the cloud (Rosenfeld and Woodley, 2000).

Because liquid or ice water content can be represented as a product of τ and r_e, if r_e is prescribed for a specific cloud scenario, water content can be parameterized by τ. To demonstrate the sensitivity to r_e and to the thermodynamic phase we create theoretical τ-Z spaces with two constant r_e for water only and then for the same r_e we compare liquid water to ice only.

In the paper we tried to mark the higher extreme of cloud cooling therefore the effective radius profile was chosen to be on the larger end with r_e=20 µm for the lower clouds and r_e=40 µm for the high and to mark the total freezing level in higher levels of the atmosphere where the temperature equals −40 °C.

Appendix B

How significant is the underestimation of high clouds effect from observations?

As mentioned in the paper, research of cloud aerosol interactions from observations requires information on both clouds and aerosols over the same area. As the cloud fraction increases the likelihood to have reliable aerosol retrieval from satellite sensors decreases. Figure 1c in the paper shows clear correlations between cloud fraction and cloud height. Higher clouds have larger coverage and therefore cloud aerosol pixels are biased toward lower clouds.

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Acknowledgements. This research was supported in part by the Minerva Foundation, the Weizmann-Argentina Cooperation program, the Yeda-Sela program and the NASA's Interdisciplinary Sciences Program under the direction of Hal Maring. I.K. is incumbent of the Benjamin H. Swig and Jack D. Weiler career development chair.

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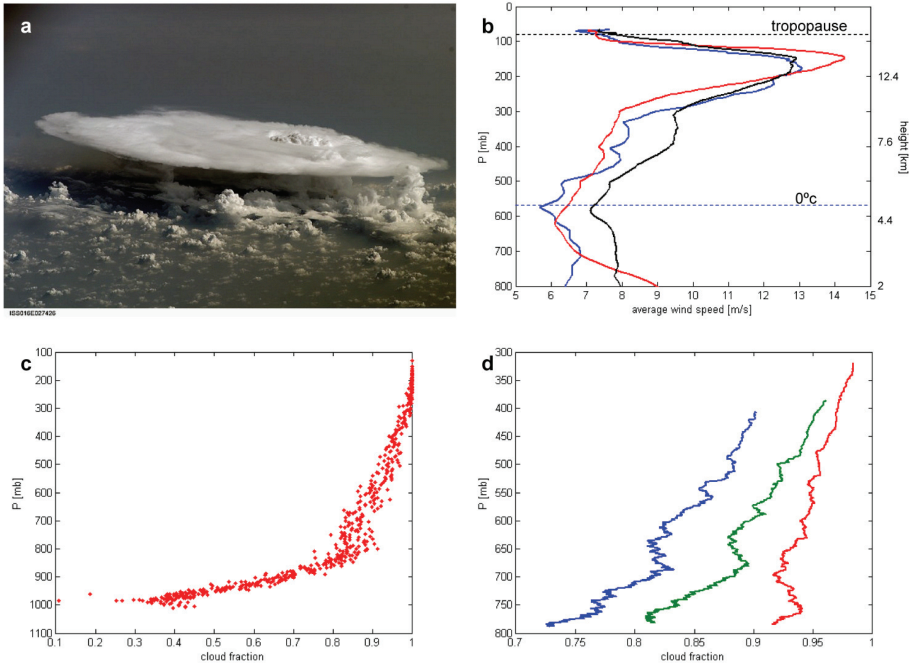

Fig. 1. **(a)** Photograph of a deep convective cloud with large anvil over Africa – 2 May 2008, photographed from the space shuttle (image courtesy of the Image Science & Analysis Laboratory, NASA Johnson Space Center). **(b)** Monthly average daytime wind profiles measured with radiosondes at 3 stations along the tropical Atlantic: Natal, Br (5.9° S 35.2° W); Ilha Fernando de Noronha, Br (3.8° S 32.4° W), Ascension Island (7.9° S 14.4° W). **(c)** Analysis of cloud fraction as a function of the cloud top pressure (height) over the ITCZ (tropical Atlantic). **(d)** Focusing on the free atmosphere (>800 hPa), cloud fraction as a function of the cloud top pressure for 3 averaged aerosol groups, AOD=0.13 (blue), 0.24 (green) and 0.41 (red).

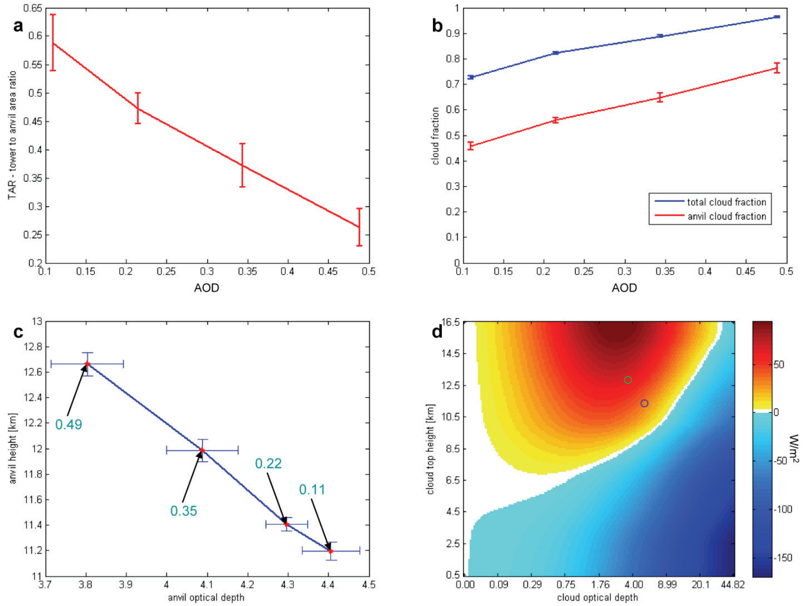

Fig. 2. The effect of aerosol on deep convective cloud tops. **(a)** Tower to anvil area ratio and **(b)** Ice cloud fraction as a function of the aerosol loading, all data (blue) and anvil fraction (red). **(c)** The anvil optical depth vs. height for 4 levels of aerosol loadings (marked with arrows). The polluted deep convective clouds are invigorated reaching higher levels of the atmosphere where the wind speed increases dramatically (Fig. 1b). Therefore, the anvil area is expected to increase and the tower to anvil ratio and cloud optical depth is expected to decrease (Fig. 2a). **(d)** τ-Z space of the cloud radiative forcing at the top of the atmosphere. The upper cooling limit of daily average cloud radiation effect in [W/m^2] as a function of the cloud height (vertical axis) and cloud optical depth (horizontal axis) for average effective radius ranging from 20 to 40 µm in the upper atmosphere and mixed phase starts from −20 to −40 °C where full freezing occurs. Warm colors represent a net warming effect. The blue circle marks the average anvil location for the cleanest set and the green for the most polluted set, note the transition left and up.

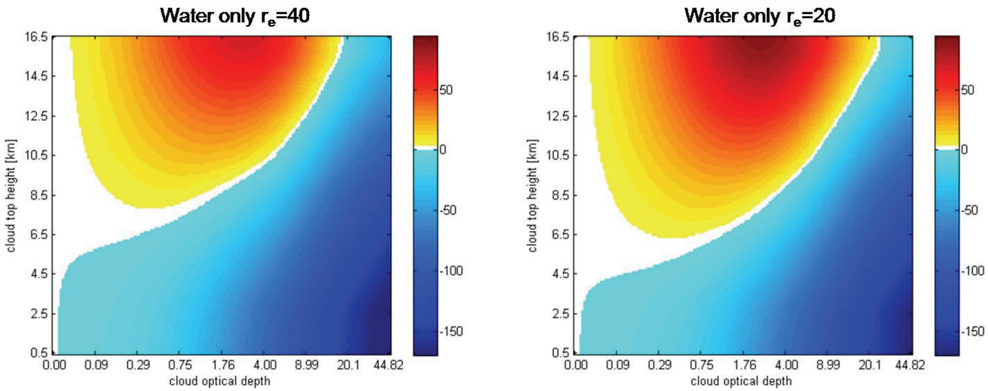

Fig. A1. τ-Z spaces for different water effective radii. Left r_e=40 µm, right r_e=20 µm. Note how the transition between cooling to warming occurs higher in the atmosphere for the large r_e. For a given cloud optical depth τ larger r_e means more cloud liquid water content, therefore the transition toward warming will occur at higher levels.

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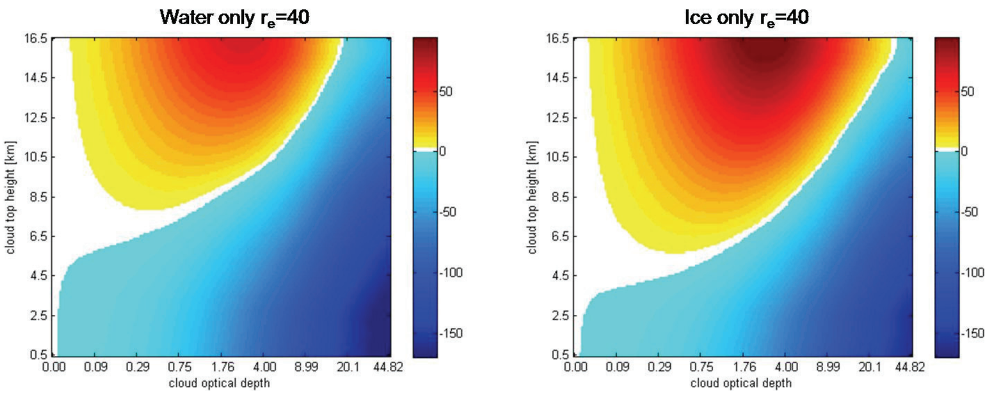

Fig. A2. τ-Z spaces for all water and all ice. Left water r_e=40 µm, right ice r_e=40 µm. Note how the transition between cooling to warming occurs higher in the atmosphere for the water cloud. Refractive index of ice is different from water and in the solar range tends to absorb more. Therefore the overall fluxes in the solar range will be smaller for ice clouds moving the transition to warming to a lower atmospheric level.

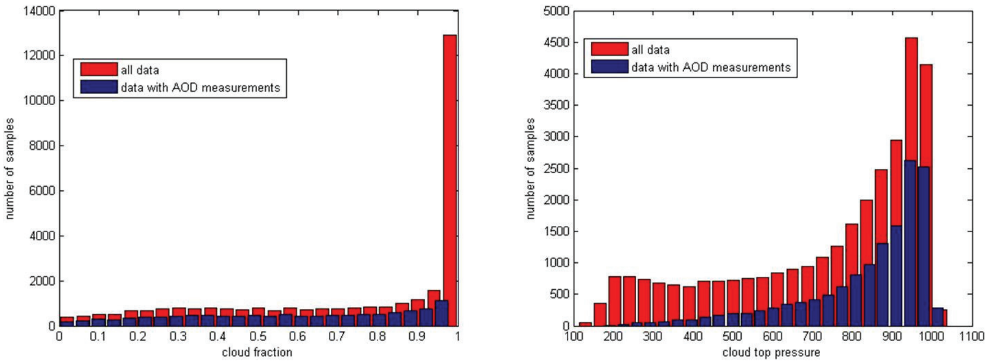

Fig. B1. Histograms of cloud fraction (left) and cloud top height (right) over the study area. Red shows all data and blue shows pixels that have both cloud and aerosol information. The study uses only the blue pixels. Note how the blue distribution misses the high clouds with the larger cloud fraction. Based on the left histogram it is easy to see that most of the effect of the high clouds (warming effect) will not be included in such cloud aerosol interaction analysis. Due to such inherent problem in observing both cloud and aerosol over the same area, it is reasonable to say that the warming effect of aerosols is highly underestimated from satellite observations.

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