Response to the reviewers' comments on

<u>Atmospheric energy budget response to idealized aerosol perturbation in tropical cloud systems</u>

We would like to thank the revisers for their constructive and thoughtful reviews that helped us improve our paper.

Below please find a point by point reply to all of the reviewers' comments (in blue).

Reviewer #1:

The authors have run some fairly large-domain cloud resolving simulations over the tropical Atlantic, in order to investigate how aerosol indirect effects contribute to the radiative impact of aerosols. They note a non-negligible contribution that depends on the cloud regime being examined. The regime and location dependence of aerosol forcing is something that is not well captured in climate models, since most don't include aerosol effects on convection. This study is straightforward and worthwhile, as it begins to break down these differences in the radiation budgets. I believe it is a good contribution and have only a few suggestions to help improve the manuscript.

<u>Reply:</u> we would like to thank the reviewer again for the effort and the constructive comments. We are happy that the reviewer found our paper to be straightforward and worthwhile.

Comments:

Changing CDNC is, you even admit, a rather simplistic way to approach aerosol effects. In addition to neglecting the activation and scavenging effects that you mention, another missing piece is the direct effect. This can be especially important in the eastern Atlantic where you are looking since a lot of the aerosols that would be present in this region are dust. Have you considered how the direct effect would fit into this?

<u>Reply:</u> Thank you for this comment. Indeed, the direct effect of aerosol is interesting and important [and is a main focus of our work, i.e. (Dagan et al., 2019)] but is not included in this study. We believe that separating the overall response of the atmospheric energy budget to the radiative and microphysical aerosol effects is a necessary first step in studying this complex system. This is the approach we are taking in the current study. However, we are currently working on implementing the aerosol model HAM (Stier et al., 2005) into the reginal version of ICON. This will allow studying the mutual interaction between the aerosol radiative and microphysical effects in a cloud and aerosol resolving simulations.

Following this comment, we have added to the revised manuscript the following:

"The different CDNC scenarios serve as a proxy for different aerosol conditions (as the first order effect of increased aerosol concentration on clouds is to increase the CDNC, Andreae, 2009). This also allows to separate the cloud response from the uncertainties involved in the representation of the aerosols in numerical models (Ghan et al., 2011; Simpson et al., 2014; Rothenberg et al., 2018). However, it limits potential feedbacks between clouds and aerosols, such as the removal of aerosol levels by precipitation scavenging and potential aerosol effects thereon. In addition, the fixed CDNC framework does not capture the differences in aerosol activation between shallow and deep clouds, due to differences in vertical velocity. Another aerosol effect that is not included in our simulations is the direct interaction between aerosol and radiation. In future work we plan to examine the mutual interaction between microphysical effects and the direct aerosol radiative effects."

In the conclusions:

"Furthermore, we do not include the temporal evolution of the aerosol concentration. Feedbacks between the aerosol concentration and clouds processes (such as wet scavenging), as well as the direct effects of aerosol on radiation would add another layer of complexity that should be accounted for in future work."

You held SST constant in these simulations. Do you have a sense for how much this might have affected the overall energy budgets? I would suspect at least the sensible heat flux might show some differences.

Reply: We agree that including interactive SST would predominantly affect the sensible heat flux. However, please note that in both cases the sensible heat flux is an order of magnitude smaller than the rest of the terms (Figs. 4 and 12 in the manuscript). Hence, we do not believe that the effect on the total energy budget would be large. In addition, due to the large heat capacity of the ocean, over two days of simulation the SST is not expected to dramatically change. For example, in the deep-cloud dominated case, the difference in the surface radiative fluxes between the clean and polluted conditions is about 12 W/m^2. Considering the heat capacity of 50m deep ocean mixed layer results in about 0.005K difference in the SST between the two cases over the two days simulation. In the shallow-cloud dominated case the difference in surface radiative fluxes in about half of the deep-cloud dominated case (6 W/m^2), resulting in half the temperature change.

Following this argument, we added a clarification about this point to the revised manuscript:

"Additional details, such as the surface and atmospheric physics parameterizations, are described in Klocke et al., (2017) and include an interactive surface flux scheme and fixed sea surface temperature (SST). We note that using a fixed SST does not include feedbacks of aerosols on the SST evolution that could change the surface fluxes. However, due to the large heat capacity of the ocean, we do not expect the SST to dramatically change over the two days simulations."

Your 'residual' term is rather large, especially in the deep convective case. You make the point earlier that this term would become negligible on longer time and spatial scales, however an important point in this paper is how large the differences can be on smaller scales. Do you have thoughts on what is largely making up this residual term? How much of it is physical processes that you are not considering, versus the fact that the model simulations are not going to be perfectly balanced, considering the scales and the boundary forcing.

Reply: What was previously called a "residual" term was changes in the revised manuscript to be refer to as "energy imbalance" as it better describes it. A recent study shows that in order to get close to energy balance the spatial scale should be on the order of ~5000km (Jakob et al., 2019) and the time scale longer than a month (we found a similar scale using GCMs). Our simulations operate on smaller spatial-temporal scales than that and hence it is not surprising that we obtain an imbalance.

The energy imbalance is composed of changes in the storage term and local divergence or convergence of dry-static energy into the domain. In our case, almost the entire imbalance is simply dry static energy that moves in or out of the domain. For example, in the deep-cloud dominated case there is a net production of dry-static energy in the domain by precipitation (which is not entirely balanced by the radiative cooling). This extra dry-static energy is then advected out of the domain.

An explanation about this point was added to the revised manuscript:

"The total column atmospheric energy budget can be described as follows:

$$LP + Q_R + Q_{SH} = div(s) + ds/dt \qquad (1)$$

Equation 1 presents a balance between the latent heating rate (LP - latent heat of condensation [L] times the surface precipitation rate [P]), the surface sensible heat flux (Q_{SH}), the atmospheric radiative heating (Q_R), the divergence of dry static energy (div(s), which will become negligible on sufficiently large spatial scales), and the dry static energy storage term (ds/dt, which will become negligible on long [inter-annual] temporal scales). Throughout the rest of this paper we *will refer to the right-hand side of Equation 1 (div(s)+ds/dt) as the energy imbalance (which is calculated as the residual [R] of the left-hand side)."*

"In this shallow-cloud dominated case the radiative cooling of the atmosphere is significantly larger than the warming due to precipitation (mean of -114.7 W/m^2 compared to 90.1 W/m^2), hence the energy imbalance (R) is negative. Negative R means that there must be some convergence of dry static energy into the domain and/or decrease in the storage term, in this case it is mostly due to convergence of dry static energy."

"Next, we analyse the atmospheric energy budget for the deep-cloud dominated case (Fiona tropical storm – Fig. 12). As opposed to the shallow-cloud dominated case, in this case the LP contribution dominates over the radiative cooling and hence the energy imbalance R is positive and large, suggesting divergence of dry static energy out of the domain."

Your mass flux in Fig 11 - how is this calculated? Is this just a total over the whole domain? Or only in updrafts? If this is domain-wide, I imagine the largest reason for the increase is simply the larger amount of deep convection.

<u>Reply:</u> Thank you for your comment that helped us clarify this point. Calculating the relative change in the cloud fraction and in the water content between clean and polluted conditions demonstrate the dominate role of the latter in the increase in mass flux. This calculation demonstrates that the cloud fraction (total water content) increases by 20% (72%) at 500mb in the simulation with CDNC=200cm^{-3} compared with the simulations with CDNC=20cm^{-3} in the deep-convection dominated case. Similarity, in the shallow dominated case the cloud fraction (total water content) increases by 22% (85%) at 500mb in the simulation with CDNC=200cm^{-3} compared with the simulations with CDNC=20cm^{-3}. These calculations demonstrate that about 80% of the increase in mass flux under polluted conditions occur due to the increase in water content and only about 20% occur due to the increase in cloud fraction (recalling that the vertical velocity is similar between the two simulations). The increase in total water content is caused by warm rain suppression at the lower troposphere.

This is now better explained in the revised manuscript:

"Both the increase in water vapor and ice content in the upper troposphere are driven by an increase in upward water (liquid and ice) mass flux with increasing CDNC (Fig. 11). An

increase in mass flux could be caused by an increase in vertical velocities and/or by an increase in cloud (or updraft) fraction and/or by an increase in cloud water content. In our case, the increases in mass flux is driven partially by the small increase in vertical velocity (especially for updraft between 5 and 10 m/s – Fig. 11), partially by the small increase in cloud faction at this level (Fig. 9) and mostly due to the larger water mass mixing ratio (Fig. 9) that leads to an increase in mass flux even for a given vertical velocity."

"Analysis of the upward water mass flux from the warm to the cold part of the clouds (at 500 mb) in the different simulations (Fig. 19), demonstrates a substantial increase with the increase in CDNC (Chen et al., 2017), which occurs due to the increase in the water content (Fig. 17) and the delay in the rain formation to higher levels (Heikenfeld et al., 2019), even without a large change in the vertical velocity or cloud fraction at this level (Fig.17)."

The paper is a bit long and I think you could consider getting rid of a few of the figures that tell a redundant story. The inclusion of the 2nd, deep convective case is important because of this point you make on page 32 "Our results demonstrate that regional atmospheric energy budgets can be significantly perturbed by changes in CDNC and that the magnitude of the effect is cloud regime dependent (even for a given geographical region and given time of the year as the two cases are separated by less than a week)." However, the physical mechanisms for the changes in cloud amount and radiative fluxes are consistent between the two cases, so some of the figures and discussion here are a bit repetitive.

Reply: Thank you for this comment. We did consider shortening the paper and not including the figures of the second case but eventually we decided to keep them in as it demonstrate an important point of this paper that the aerosol effect on the atmospheric energy budget is meteorological conditions dependent.

In general this could use some copy-editing. Nothing that prevents understanding, but there are a number of small typos and verb agreement issues.

Reply: Thank you. The manuscript went thought copy editing and corrected accordingly.

Reviewer #2:

This paper studies the differences in the radiative and energy budgets in a tropical environment when the cloud droplet number concentration (CDNC) is changed (as a proxy for changes to the environmental aerosol concentration). The simulated periods are two separate 2-day within the same week where the convection is either predominantly shallow or predominantly deep - each case is simulated with CDNC values. The authors find substantial changes in amount of energy absorbed by the atmosphere by changing the CDNC and find substantial differences in the response between the two sets of simulations.

The main component of the paper is a breakdown of the energy budget into radiative, sensible heating and heating through precipitation formation. The radiative component is later broken down into shortwave and longwave fluxes at both the surface and top of the atmosphere. The differences are also quantified in the time evolution of near- surface temperature, precipitation, cloud fraction and in-cloud water contents.

Overall, I find the study to be well formulated with a clear motivation and simple but successful strategy for breaking apart the components contributing to the changes in the atmospheric energy budget. There are no substantial shortcomings that should prevent the publication of this study; however, I have a few suggestions that could improve this contribution which are explained below.

Reply: we would like to thank the reviewer again for the effort and for the suggestions. We are happy that the reviewer found that our paper well formulated.

Primarily my suggestions are aimed to help the authors achieve their stated aim of better understanding the physical processes behind aerosol effects on the atmospheric energy budget. I see that their study does indeed achieve this, at least partly, but that these results are not clearly expressed in the abstract nor the conclusions. Throughout the paper, the authors do a good job of describing the differences between their simulations and quantifying these differences (although in parts the quantification could be improved) - however, it is mostly left to the reader to put these pieces of information together to get an understanding of the physical processes involved. As a result, my overall impression of the authors conclusions and abstract are: "we found another case where aerosol-cloud interactions behave differently under different environmental conditions," which could be relatively simply converted to "these processes (*see below) contribute to the different energy budget changes for shallow and deep convection when CDNC is changed"

(1) From my understanding of the presented results, it seems that the large difference between the shallow case and the deep case is the potential for a large upper-level cloud fraction change in the deep case. I understand this as an increase of the anvil area, resulting in reduced LW emission from the surface/lower atmosphere and therefore a warming contribution of the larger anvil. In the shallow case, the upper level cloud fraction also has a systematic change, but because it occupies a smaller part of the model domain - the overall change in the energy budget is controlled by the change in low cloud fraction and the Twomey effect. If the authors agree with this, I suggest adding a paragraph into the conclusions and a sentence in the abstract clarifying these physical changes in the model and their impact on the energy budget.

Reply: Thank you for this suggestion that help us clarify this point. The reviewer's point is a main conclusion of our paper; however, we believe that our results present more than just that and include (among other) the effect of the thermodynamic evolution on radiation under different CDNC conditions, the effect of CDNC on surface fluxes and so on. Nevertheless, following the reviewer's comment, we have added a clarification to the revised manuscript.

In the abstract:

"It is shown that the total column atmospheric radiative cooling is substantially reduced with CDNC in the deep-cloud dominated case (by ~10.0 W/m^2), while a much smaller reduction (~1.6 W/m^2) is shown in the shallow-cloud dominated case. This trend is caused by an increase in the ice and water vapor content at the upper troposphere that leads to a reduced outgoing longwave radiation, an effect which is stronger under deep-cloud dominated conditions."

In the conclusions section:

"Both the increase in water vapor and ice content in the upper troposphere are driven by an increase in water mass flux with increasing CDNC to these levels (Fig. 19, (Koren et al., 2005; Rosenfeld et al., 2008; Altaratz et al., 2014; Chen et al., 2017)), which is caused mostly by the increase in the water mixing ratio in the mid-troposphere rather than by increase in vertical velocity (Fig. 19) or in cloud fraction (Fig. 17). The ice content in the upper troposphere is also increased due to reduction in the ice falling speed (Grabowski and Morrison, 2016), while the increased relative humidity at these levels, further increases the ice particle lifetime due to slower evaporation."

"In the shallow-cloud dominated case (which also contains a significant amount of deep convection), the response of Q_R is weaker but still substantial (a total decrease in the atmospheric radiative cooling of 1.6 W/m^2 - Fig. 20). The weaker total response under the shallow-cloud dominated conditions is due to the smaller role of the ice part in this case."

(2) Breakdown of the vertical mass flux changes with CDNC into component parts

The vertical mass flux of water is shown to change between the simulations with different CDCN. What is the cause for this change? Either the vertical velocity should be increasing (which seems not to be the case from the vertical velocity distributions in Figures 11 and 19), so either the updraft area is increasing [implying wider updrafts?] or the in-cloud water mass in increasing [because of a less efficient precipitation-forming processes?]. To what extent are these two factors important? Furthermore, what happens to the vertical mass flux at (e.g.) 800 hPa - where the total water content is quite similar between all CDNC concentrations - is there still an increased vertical mass flux?

Reply: Thank you for this comment that helped us clarify this point. Calculating the relative change in the cloud fraction and in the water content between clean and polluted conditions demonstrate the dominate role of the later in the increase in mass flux. This calculation demonstrates that the cloud fraction (total water content) increases by 20% (72%) at 500mb in the simulation with CDNC=200cm^{-3} compared with the simulations with CDNC=20cm^{-3} in the deep convection dominated case. Similarity, in the shallow dominated case the cloud fraction (total water content) increases by 22% (85%) at 500mb in the simulation with CDNC=200cm^{-3} compared with the simulations with CDNC=20cm^{-3}. These calculations demonstrate that about 80% of the increase in mass flux under polluted conditions occur due to the increase in water content and only about 20% occur due to the increase in cloud fraction (recalling that the vertical velocity is similar between the two simulations). The increase in total water content is caused by warm rain suppuration at the lower troposphere.

This is now better explained in the revised manuscript:

"Both the increase in water vapor and ice content in the upper troposphere are driven by an increase in upward water (liquid and ice) mass flux with increasing CDNC (Fig. 11). An increase in mass flux could be caused by an increase in vertical velocities and/or by an increase in cloud (or updraft) fraction and/or by an increase in cloud water content. In our case, the increases in mass flux is driven partially by the small increase in vertical velocity (especially for updraft between 5 and 10 m/s – Fig. 11), partially by the small increase in cloud faction at this level (Fig. 9) and mostly due to the larger water mass mixing ratio (Fig. 9) that leads to an increase in mass flux even for a given vertical velocity."

"Analysis of the upward water mass flux from the warm to the cold part of the clouds (at 500 mb) in the different simulations (Fig. 19), demonstrates a substantial increase with the increase

in CDNC (Chen et al., 2017), which occur due to the increase in the water content (Fig. 17) and the delay in the rain formation to higher levels (Heikenfeld et al., 2019), even without a large change in the vertical velocity or cloud fraction at this level (Fig.17).”

In addition, we calculated the mass flux at the level of 800 mb as the reviewer suggested (Figs. R1 and R2 below). It demonstrates that in this level there is also a general increase in mass flux with the increase in CDNC but to a lesser extent. The increase in mass flux is driven by a small increase in water content (Figs. 9 and 17) and a small increase in vertical velocity (Figs. R1 and R2), while the cloud fraction is similar between the simulations (Figs. 9 and 17). In the manuscript we still present the mass flux at 500 mb as it represents the transition between the warm and the cold parts of the clouds.

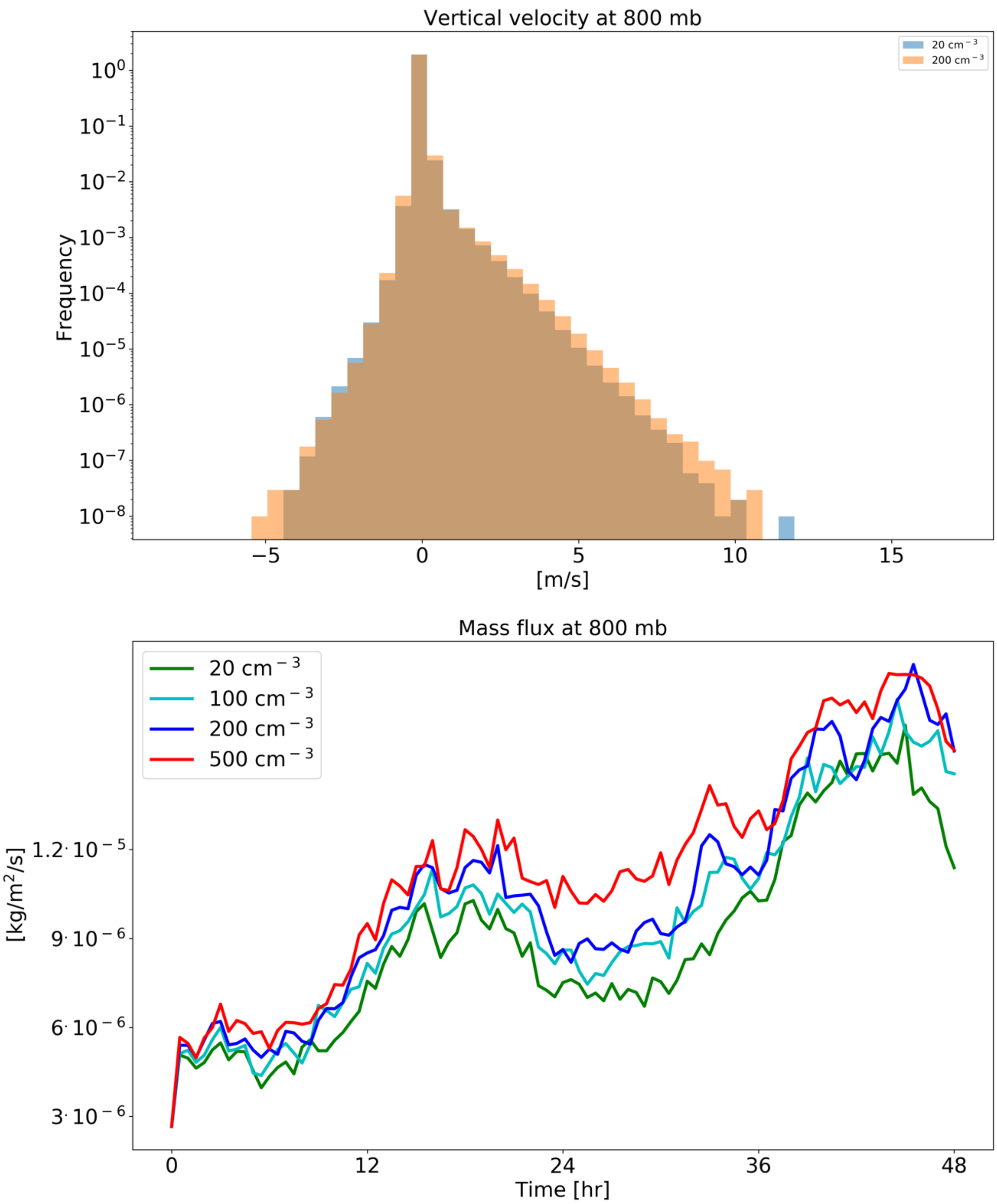


Figure R1. histograms of ICON simulated vertical velocity at the level of 800 mb for a clean (CDNC = 20 cm^{-3}) and polluted (CDNC = 200 cm^{-3}) simulations (upper), and the time evolution of the net upwards water (liquid and ice) mass flux (lower) for the different CDNC simulations for the shallow-cloud dominated case (10-12/08/2016). In the histogram only two simulations are presented for clarity.

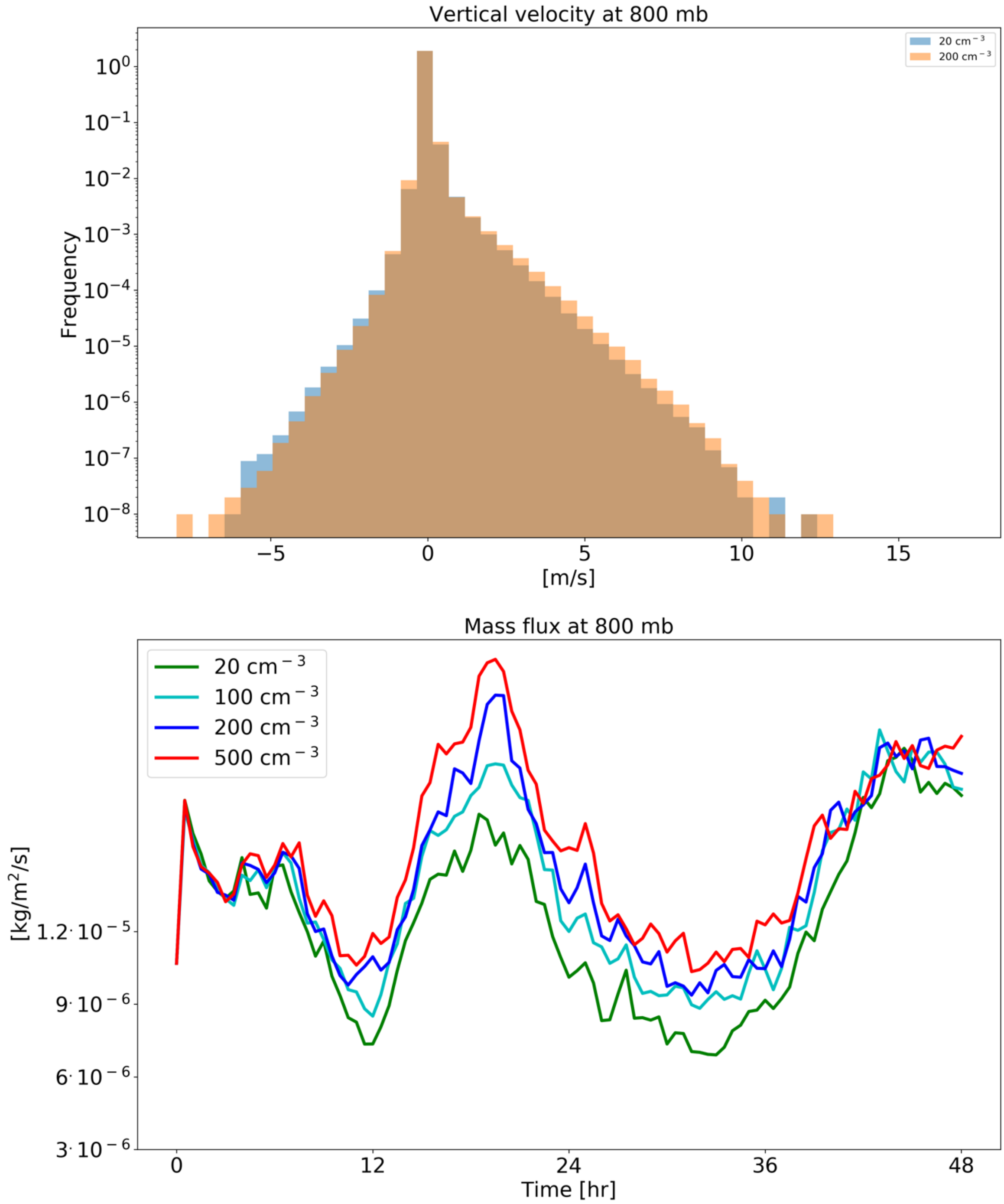


Figure R2. histograms of ICON simulated vertical velocity at the level of 800 mb for a clean (CDNC = 20 cm^{-3}) and polluted (CDNC = 200 cm^{-3}) simulations (upper), and the time evolution of the net upwards water (liquid and ice) mass flux (lower) for the different CDNC simulations for the deep-cloud dominated case (16-18/08/2016). In the histogram only two simulations are presented for clarity.

(3) Large contribution from residuals

A large contribution to the overall energy budget is within the residual term, which the authors state should reduce to zero given a long enough averaging time. How can the authors be sure that this is true and that the large component in the residual term is not a "buffering" effect - e.g. changing stability of the atmosphere to compensate for the changing energy budget? Can the 3D distribution of the residual values be used to quantify this at all?

Reply: A similar argument was raised by reviewer #1 – we will repeat our reply here. What was previously called a "residual" term was changes in the revised manuscript to be refer to as "energy imbalance" as it better describes it. A recent study shows that in order to get close to energy balance the spatial scale should be on the order of ~5000km (Jakob et al., 2019) and the time scale longer than a month (we found a similar scale using GCMs). Our simulations operate on smaller spatial-temporal scales than that and hence it is not surprising that we obtain an imbalance.

The energy imbalance is composed of changes in the storage term and local divergence or convergence of dry-static energy into the domain. In our case, almost the entire imbalance is simply dry static energy that moves in or out of the domain. For example, in the deep-cloud dominated case there is a net production of dry-static energy in the domain by precipitation (which is not entirely balanced by the radiative cooling). This extra dry-static energy is then advected out of the domain.

An explanation about this point was added to the revised manuscript:

"The total column atmospheric energy budget can be described as follows:

$$LP + Q_R + Q_{SH} = div(s) + ds/dt \qquad (1)$$

Equation 1 presents a balance between the latent heating rate (LP - latent heat of condensation [L] times the surface precipitation rate [P]), the surface sensible heat flux (Q_{SH}), the atmospheric radiative heating (Q_R), the divergence of dry static energy (div(s), which will become negligible on sufficiently large spatial scales), and the dry static energy storage term (ds/dt, which will become negligible on long [inter-annual] temporal scales). Throughout the rest of this paper we will refer to the right-hand side of Equation 1 (div(s)+ds/dt) as the energy imbalance (which is calculated as the residual [R] of the left-hand side)."

"In this shallow-cloud dominated case the radiative cooling of the atmosphere is significantly larger than the warming due to precipitation (mean of -114.7 W/m^2 compared to 90.1 W/m^2), hence the energy imbalance (R) is negative. Negative R means that there must be some convergence of dry static energy into the domain and/or decrease in the storage term, in this case it is mostly due to convergence of dry static energy."

"Next, we analyse the atmospheric energy budget for the deep-cloud dominated case (Fiona tropical storm – Fig. 12). As opposed to the shallow-cloud dominated case, in this case the LP contribution dominates over the radiative cooling and hence the energy imbalance R is positive and large, suggesting divergence of dry static energy out of the domain."

(4) There appears to be a mismatch between TWP in Figure 8 (lower right plot) and qt in Figure 9 (lower central plot). Similarly in Figures 16 & 17. The vertical profile of qt is quite similar for 3 simulations in the shallow case (Figure 9; excluding the 500 cm-3 line). Similarly, the qt values from 3 simulations in the deep case are also similar (Figure 17; excluding the 20 cm-3 line). However, in figure 8 & 16 these is clear separation between all the TWP lines throughout the simulation. By quick calculation the spread in the TWP timeseries seems too large to be explained by the differences in qt (which are mostly between 650-400 hPa). How can this difference be explained? Is the TWP only including cloud and ice, but ignoring rain water? Similarly, is the LW only cloud, ignoring rain?

<u>Reply:</u> Thank you for this comment that helped us clarify this point. Indeed, the LWP include the cloud mass (qc) and not the rain mass (qr). This is done for consistency with LWP calculated from satellite observations, which are sensitive only to the cloud mass and not to the rain mass (see also the reply to the next point). This is now better explained in the revised manuscript:

"Figure 8. Domain average properties as a function of time for the different CDNC simulations for the shallow-cloud dominated case. The properties that are presented here are: cloud fraction (CF), rain rate, temperature in 2 m, liquid water path (LWP – based on the cloud water mass, excluding the rain mass for consistency with satellite observations), ice water path (IWP) and total water path (TPW = LWP + IWP). For each property, the mean difference between all combinations of simulations, normalized to a factor 5 increase in CDNC, and its standard deviation appear in parenthesis."

(5) Following from the above point: is rain water radiatively interactive in the model? If not, to what extent does this removal of mass from the radiatively interactive cloud species have on the Twomey effect calculations performed, given that the rain water mass is almost equal to the cloud water mass at some heights?

<u>Reply:</u> As done in most atmospheric models (Hill et al., 2018), the rain mass is not included in the radiative calculations. Since the rain drops are much larger than the cloud droplet, their cross-section available for interaction with radiation (for a given water mass) is much smaller

and usually negligible (Hill et al., 2018). For example, a simple "back of the envelop" calculation of the cloud optical depth (τ) as in Heus & Seifert (2013) and Spill et al. (2019) follows as:

$\tau = 0.19 * LWP^{5/6} * N^{1/3}$,

where LWP is the liquid water path and N is the drop concentration, for a given LWP for cloud droplets with radius r=10 µm and for rain drops with r=0.5 mm yield a factor of 50 decrease in τ for the rain compare with the cloud. Generally, as τ is proportional to $N^{1/3}$ it decreases proportional to r (for a given LWP), and since the cloud droplets and rain drops are separated by 1 or 2 orders of magnitudes, the effect of the cloud droplets on the radiation is much larger than that of the rain drops. This simple calculation does not account for the changes in Mie size parameter between rain drops and cloud droplets but it serves to demonstrate the orders of magnitude differences between the two different regimes.

(6) Impact of simplifications

The approach of simply modifying the CDNC instead of the aerosol concentration of the atmosphere ignores several potentially important processes/feedbacks (e.g. activation of CCN/IN, size distribution of aerosol, direct radiative effects) - the authors should comment on these shortcomings in the conclusions.

Reply: Thank you. Based on this comment we have added a comment about the limitation of using a fixed CDNC simulations:

"The different CDNC scenarios serve as a proxy for different aerosol conditions (as the first order effect of increased aerosol concentration on clouds is to increase the CDNC, Andreae, 2009). This also allows to separate the cloud response from the uncertainties involved in the representation of the aerosols in numerical models (Ghan et al., 2011; Simpson et al., 2014; Rothenberg et al., 2018). However, it limits potential feedbacks between clouds and aerosols, such as the removal of aerosol levels by precipitation scavenging and potential aerosol effects thereon. In addition, the fixed CDNC framework does not capture the differences in aerosol activation between shallow and deep clouds, due to differences in vertical velocity. Another aerosol effect that is not included in our simulations is the direct interaction between aerosol and radiation. In future work we plan to examine the mutual interaction between microphysical effects and the direct aerosol radiative effects."

In the conclusions:

"Furthermore, we do not include the temporal evolution of the aerosol concentration. Feedbacks between the aerosol concentration and clouds processes (such as wet scavenging), as well as the direct effects of aerosol on radiation would add another layer of complexity that should be accounted for in future work."

(7) Robustness of results

The authors should comment on the robustness of these results, in light of the fact that single simulations (rather than ensembles) of two individual case studies are performed. The results in figures 8 & 16 suggest a clear separation between all 4 CDNC concentrations from early on in the simulation - however, the vertical profiles of qi, qt and CF in figures 9 & 17 suggest that the 20 CDNC cm-3 simulation is the only one of the four that is substantially different (particularly at upper levels, which seem to be most important in this story).

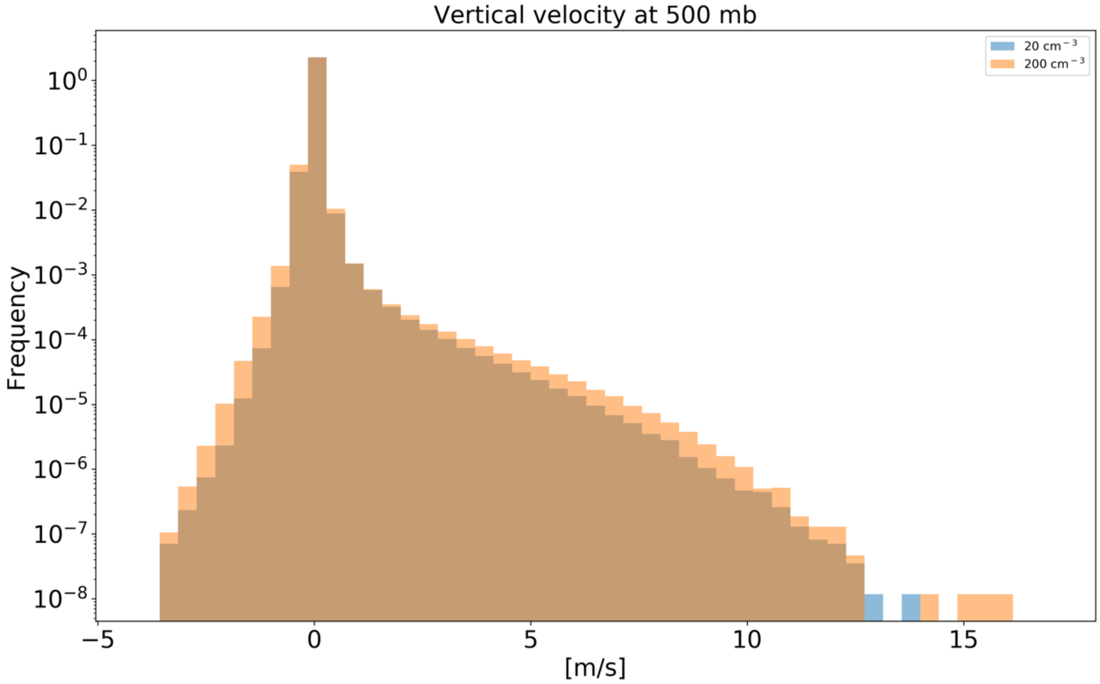
Reply: The robustness of our simulations, which, as the reviewer stated, are based on few simulations rather then on large ensemble, occupied our mind as well. In a recent paper which is currently under discussion in ACPD (Dagan and Stier, 2019) we investigate this exact question. In that paper we use a smaller domain compared to the current study (3^{o} x 3^{o} rather than 22^{o} x 11^{o}) to simulated a large ensemble of initial conditions (all together we simulate 124 different simulations). This large ensemble enables robust identification of the effect of CDNC changes on cloud properties. We were able to show that the general conclusions which are stated in the current study hold also for large statistics. Based on the following comment we have added a discussion about the robustness of our results to the revised manuscript:

"There exists a large spread in estimates of aerosol effects on clouds for different cloud types and different environmental conditions. In this study, as we use a relatively large domain (22^{o} x 11^{o}) and two different dates (each for two days), we sample many different local environmental conditions and cloud types. Such more realistic setups (although with lower spatial resolution) could provide more reliable estimates of aerosol effects on heterogeneous cloud systems than just one-cloud-type, small domain simulations (as was done in many previous studies, e.g (Dagan et al., 2017; Seifert et al., 2015; Ovchinnikov et al., 2014)). However, the conclusions demonstrated here are based on two specific cases. In order to examine the validity of our main conclusions over a wider range of initial conditions, we have conducted a large ensemble of simulations starting from realistic initial conditions (although with a smaller domain) in a companion paper (Dagan and Stier, 2019). These simulations

demonstrate that the main conclusions presented in this paper are robust and hold also for a wide range of initial conditions representative for this area."

Minor points: please show the mass flux from all four simulations in figures 11 and 19 to be consistent with the other plots in the paper.

Reply: The mass flux of all four simulations are now presented in Figs. 11 and 19:



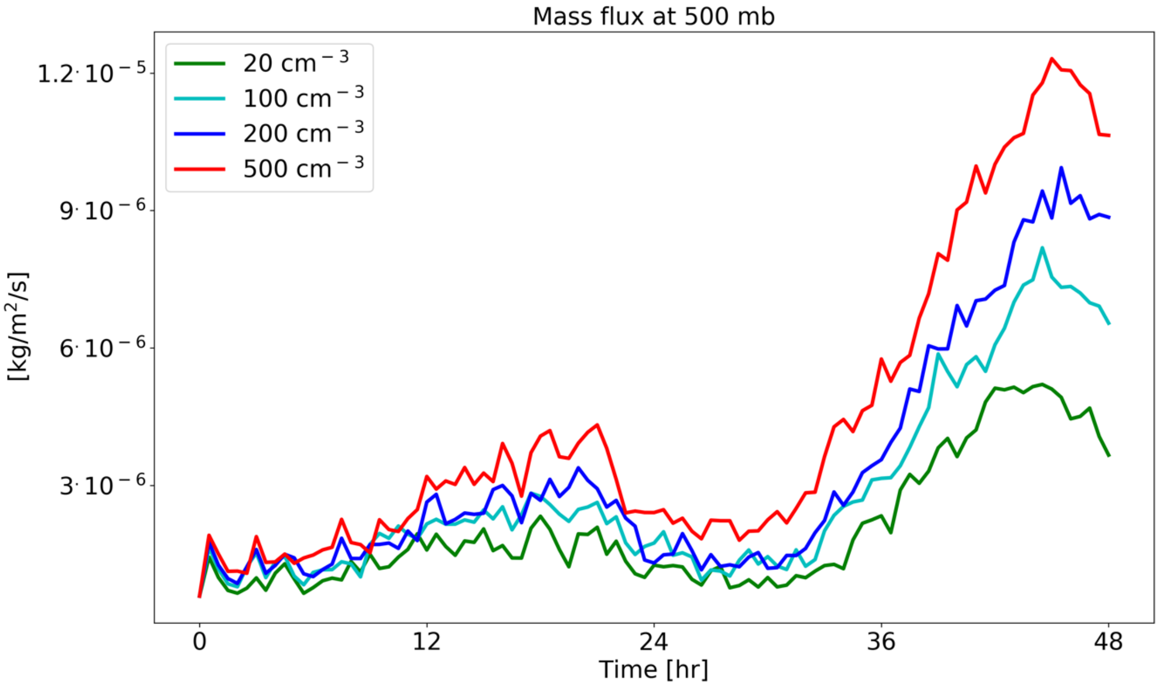


Figure 11. histograms of ICON simulated vertical velocity at the level of 500 mb for a clean (CDNC = 20 cm^{-3}) and polluted (CDNC = 200 cm^{-3}) simulations (upper), and the time evolution of the net upwards water (liquid and ice) mass flux (lower) for the different CDNC simulations for the shallow-cloud dominated case (10-12/08/2016). The 500 mb level is chosen as it represents the transition between the warm part to the cold part of the clouds. In the histogram only two simulations are presented for clarity.

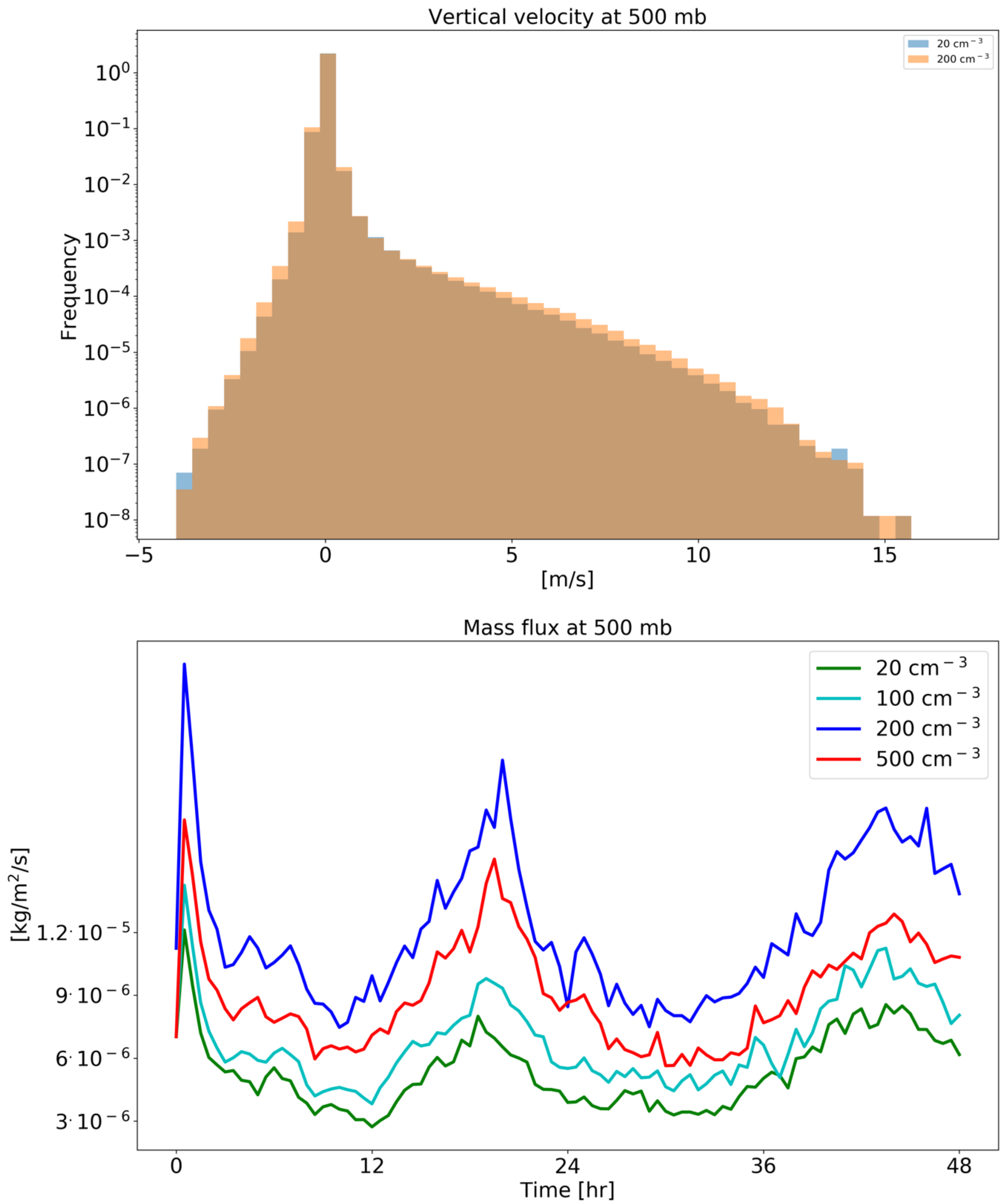


Figure 19. histograms of ICON simulated vertical velocity at the level of 500 mb for a clean (CDNC = 20 cm^{-3}) and polluted (CDNC = 200 cm^{-3}) simulations (upper), and the time evolution of the net upwards water (liquid and ice) mass flux (lower) for the different CDNC simulations for the deep-cloud dominated case (16-18/08/2016). The 500 mb level is chosen as it represents the transition between the warm part to the cold part of the clouds. In the histogram only two simulations are presented for clarity.

Lines 335-338: please be more quantitative about the results of the test with the offline radiation calculations as to the relative contributions of the cloud fraction and TWP changes. Line 367: please quantify the "vast majority" of LW flux changes due to cloudy rather than clear skies.

Reply: Thank you. More information was added to the revised manuscript:

"For estimating the relative contribution of the changes in CF and water content to the SW flux changes we have conducted off-line radiative transfer sensitivity tests. To quantify the water content radiative effect, we feed the same CF vertical profile from the model into the offline radiative transfer model BUGSrad, while allowing the water content vertical profile to change (and visa versa to compute the CF radiative effect). This approach demonstrates that the contribution from the small reduction in CF is negligible compared to the increased SW reflectance caused by the increased water content (the effect of the reduction in CF compensate only about 1% of the effect of the increase in the water content)."

"The increased humidity at the upper troposphere would act to decrease the outgoing LW flux, similar to the effect of the increased ice content in the upper troposphere (Fig. 9). However, sensitivity studies with off-line radiative transfer calculations using BUGSrad demonstrate that the vast majority (more than 99%) of the different in $\boldsymbol{F_{LW}^{TOA}}$ *between clean and polluted conditions emerges from the cloudy skies (rather than clear-sky), suggesting that the effect of the increased ice content at the upper troposphere dominates."*

The plots in figures 10 and 18, currently described in the caption as Hovmöller plots, would be better described as time-height plots.

Reply: Thanks. It was changed according to the reviewer's suggestion.

Is there an explanation for the relative minimum of cloud water content at 650 hPa in all simulations? I struggle to find a physical explanation for this.

Reply: We think that the relative minimum of domain-wide cloud water at 650 hPa is simply due to the fact that below this level there are still quite a lot of shallow clouds while above it there are anvils clouds with longer lifetime (and hence the larger cloud water mass) then the liquid at lower levels.

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Atmospheric energy budget response to idealized aerosol perturbation in tropical cloud systems

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Abstract

The atmospheric energy budget is analysed in numerical simulations of tropical cloud systems~~.~~ ~~This is done in order~~ to better understand the physical processes behind aerosol effects on the atmospheric energy budget. The simulations include both shallow convective clouds and deep convective tropical clouds over the Atlantic Ocean. Two different sets of simulations, at different dates (10-12/8/2016 and 16-18/8/2016), are ~~being~~ simulated with different dominant cloud modes (shallow or deep). For each case, the cloud droplet number concentrations (CDNC) is varied as a proxy for changes in aerosol concentrations. It is shown that the total column atmospheric radiative cooling is substantially reduced with CDNC in the deep-cloud dominated case (by ~10.0 W/m^2), while a much smaller reduction (~1.6 W/m^2) is shown in the shallow-cloud dominated case. This trend is caused by an increase in the ice and water vapor content at the upper troposphere that leads to a reduced outgoing longwave radiation, an effect which is stronger under deep-cloud dominated conditions. A decrease in sensible heat flux (driven by increase in the near surface air temperature) reduces the warming by ~1.4 W/m^2 in both cases. It is also shown that the cloud fraction response behaves in opposite ways to an increase in CDNC, showing an increase in the deep-cloud dominated case and a decrease in the shallow-cloud dominated case. This demonstrates that under different environmental conditions the response to aerosol perturbation could be different.

Introduction

The negative anthropogenic radiative forcing due to aerosols is acting to cool the climate and to compensate some of the warming due to increase in greenhouse gases (Boucher et al., 2013). However, quantification of this effect is highly uncertain with a revised uncertainty range of −1.60 to −0.65 W/m^2 (Bellouin et al., 2019). The total anthropogenic aerosol radiative forcing is composed of contribution from direct interaction of aerosols with radiation (scattering and absorption) and from indirect interaction with radiation due to changes in cloud properties.

Beside its effect on the radiation budget, aerosols may affect the precipitation distribution and total amount (Levin and Cotton, 2009; Albrecht, 1989; Tao et al., 2012). A useful perspective to improve our understanding of aerosol effect on precipitation, which became common in the last few years, arises from constraints on the energy budget (O'Gorman et al., 2012; Muller and O'Gorman, 2011; Hodnebrog et al., 2016; Samset et al., 2016; Myhre et al., 2017; Liu et al., 2018; Richardson et al., 2018; Dagan et al., 2019a). On long time scales, any precipitation perturbations by aerosol effects will have to be balanced by changes in radiation fluxes, sensible heat flux or by divergence of dry static energy. The energy budget constraint perspective was found useful to explain both global (e.g. (Richardson et al., 2018)) and regional (Liu et al., 2018; Dagan et al., 2019a) precipitation response to aerosol perturbations in global scale simulations. In this study, we investigate the energy budget response to aerosol perturbation on a regional scale using high resolution cloud resolving simulations. This enables an improved understanding of the microphysical processes controlling atmospheric energy budget perturbations. The strong connection between the atmospheric energy budget and convection has long been appreciated (e.g. (Arakawa and Schubert, 1974; Manabe and Strickler, 1964)) as well as the connection to the general circulation of the atmosphere (Emanuel et al., 1994).

The total column atmospheric energy budget can be described as follows:

$$LP + Q_R + Q_{SH} = \mathrm{div}(s) + \mathrm{d}s/\mathrm{dt} \qquad (1)$$

Equation 1 presents a balance between the latent heating rate (LP - latent heat of condensation [L] times the surface precipitation rate [P]), the surface sensible heat flux (Q_{SH}), the atmospheric radiative heating (Q_R), the divergence of dry static energy (div(s), which will become negligible on sufficiently large spatial scales), and the dry static energy storage term (ds/dt, which will become negligible on long [inter-annual] temporal scales). Throughout the rest of this paper we will refer to the right-hand side of Equation 1 (div(s)+ds/dt) as the energy imbalance (which is calculated as the residual [R] of the left-hand side).

Q_R is defined as:

$$Q_R = (F_{SW}^{TOA} - F_{SW}^{SFC}) + (F_{LW}^{TOA} - F_{LW}^{SFC}) \qquad (2)$$

and represents the rate of net atmospheric diabatic warming due to radiative shortwave (SW) and longwave (LW) fluxes. It is expressed by the sum of the surface (SFC) and top of the atmosphere (TOA) fluxes, when all fluxes are positive downwards. As in the case of TOA radiative forcing, aerosols could modify the atmospheric energy budget by both direct interaction with radiation and by microphysical effects on clouds. The latter is the focus of this study.

The microphysical effects are driven by the fact that aerosols serve as cloud condensation nuclei (CCN) and ice nuclei (IN). Larger aerosol concentrations, e.g. by anthropogenic emissions, could lead to larger cloud droplet and ice particle concentrations (Andreae et al., 2004; Twomey, 1977; Hoose and Möhler, 2012). Changes in hydrometer concentration and size distribution were shown to affect clouds' microphysical processes rates (such as condensation, evaporation, freezing and collision-coalescence), which in turn could affect the dynamics of the clouds (Khain et al., 2005; Koren et al., 2005; Heikenfeld et al., 2019; Chen et al., 2017;Altaratz et al., 2014; Seifert and Beheng, 2006a), the rain production (Levin and Cotton, 2009; Albrecht, 1989; Tao et al., 2012) and the clouds' radiative effect (Koren et al., 2010; Storelvmo et al., 2011; Twomey, 1977; Albrecht, 1989). The aerosol effect, and in particular its effects on the radiation budget and the atmospheric energy budget, is cloud regime dependent (Altaratz et al., 2014; Lee et al., 2009; Mülmenstädt and Feingold, 2018; van den Heever et al., 2011; Rosenfeld et al., 2013; Glassmeier and Lohmann, 2016; Gryspeerdt and Stier, 2012; Christensen et al., 2016), time dependent (Dagan et al., 2017; Gryspeerdt et al., 2015; Seifert et al., 2015; Lee et al., 2012; Dagan et al., 2018c), aerosol type and size distribution dependent (Jiang et al., 2018; Lohmann and Hoose, 2009) and (even for a given cloud regime) meteorological conditions dependent (Dagan et al., 2015a; Fan et al., 2009; Fan et al., 2007; Kalina et al., 2014; Khain et al., 2008) and was shown to be non-monotonic (Dagan et al., 2015b; Jeon et al., 2018; Gryspeerdt et al., 2019; Liu et al., 2019). Hence the quantification of the global mean radiative effect is extremely challenging (e.g. (Stevens and Feingold, 2009; Bellouin et al., 2019)).

Previous studies demonstrated that the mean aerosol effect on deep convective clouds can increase the upward motion of water, and hence also increase the cloud anvil mass and extent (Fan et al., 2010; Chen et al., 2017; Fan et al., 2013; Grabowski and Morrison, 2016). The increase in mass flux to upper levels was explained by the convective invigoration hypothesis (Fan et al., 2013; Koren et al., 2005; Rosenfeld et al., 2008; Seifert and Beheng, 2006a; Yuan et al., 2011a; Williams et al., 2002), which was proposed to lead to stronger latent heat release under higher aerosol concentrations and hence stronger vertical velocities. In addition to the stronger vertical velocities, under polluted conditions the smaller hydrometers are being transported higher in the atmosphere (for a given vertical velocity (Chen et al., 2017; Koren et al., 2015; Dagan et al., 2018a)) and their lifetime at the upper troposphere is longer (Fan et al., 2013; Grabowski and Morrison, 2016). The invigoration mechanism can also lead to an increase in precipitation (Khain, 2009; Altaratz et al., 2014). Both the increase in precipitation and the increase in anvil coverage would act to warm the atmospheric column: the increased precipitation by latent heat release, and the increased anvil mass and extent by longwave radiative warming (Koren et al., 2010; Storelvmo et al., 2011). However, it should be pointed out that the uncertainty underlying these proposed effects remain significant (White et al., 2017; Varble, 2018). In addition, aerosol effects on precipitation from deep convective cloud was shown to be non-monotonic and depend on the aerosol range (Liu et al., 2019).

In the case of shallow clouds, aerosol effect on precipitation was also shown to be non-monotonic (Dagan et al., 2015a; Dagan et al., 2017). However, unlike in the deep clouds case, the mean effect on precipitation, under typical modern-day conditions, is thought to be negative (Albrecht, 1989; Rosenfeld, 2000; Jiang et al., 2006; Xue and Feingold, 2006; Dagan and Chemke, 2016). The aerosol effect on shallow cloud cover and mean water mass (measure by liquid water path - LWP) might also depend on the meteorological conditions and aerosol range (Dagan et al., 2015b; Dagan et al., 2017; Gryspeerdt et al., 2019; Dey et al., 2011; Savane et al., 2015) and is the outcome of competition between different opposing response of: rain suppression (that could lead to increase in cloud lifetime and coverage (Albrecht, 1989)), warm clouds invigoration (that could also lead to increase in cloud coverage and LWP (Koren et al., 2014; Kaufman et al., 2005; Yuan et al., 2011b)) and increase in entrainment and evaporation (that could lead to decrease in cloud coverage (Small et al., 2009; Jiang et al., 2006; Costantino and Bréon, 2013; Seigel, 2014)). Another addition to this complex response is the fact that the aerosol effect on warm convective clouds was shown to be time dependent and affected by the clouds' feedbacks on the thermodynamic conditions (Seifert et al., 2015; Dagan et al., 2016; Dagan et al., 2017; Lee et al., 2012; Stevens and Feingold, 2009; Dagan et al., 2018b). Previous simulations that contained several tropical cloud modes demonstrate that increase in aerosol concentrations can lead to suppression of the shallow mode and invigoration of the deep mode (van den Heever et al., 2011). Hence the domain mean effect, even if it is demonstrated to be small, may be the result of opposing relatively large contributions from the different cloud modes (van den Heever et al., 2011). The small domain mean effect may suggest that on large enough scales the energy (Muller and O'Gorman, 2011; Myhre et al., 2017) or water budget (Dagan et al., 2019b) constrain precipitation changes.

Previous studies, using global simulations (O'Gorman et al., 2012; Muller and O'Gorman, 2011; Hodnebrog et al., 2016; Samset et al., 2016; Myhre et al., 2017; Liu et al., 2018; Richardson et al., 2018; Dagan et al., 2019a), demonstrated the usefulness of the atmospheric energy budget perspective in constraining aerosol effect on precipitation. However, the physical processes behind aerosol-cloud microphysical effects on the energy budget are still far from being fully understood. In this study we use cloud resolving simulations to increase our understanding of the effect of microphysical aerosol-cloud interactions on the atmospheric energy budget.

Methodology

The icosahedral nonhydrostatic (ICON) atmospheric model (Zängl et al., 2015) is used in a limited area configuration. ICON's non-hydrostatic dynamical core was evaluated with several idealized cases (Zängl et al., 2015). The simulations are conducted such that they are aligned with the NARVAL 2 (Next-generation Aircraft Remote-Sensing for Validation Studies (Klepp et al., 2014; Stevens et al., 2019; Stevens et al., 2016)) campaign, which took place during August 2016 in the western part of the northern tropical Atlantic. We use existing NARVAL 2 convection-permitting simulations (Klocke et al., 2017) as initial and boundary conditions for our simulations.

The domain covers ~22^{o} in the zonal direction (25^{o} - 47^{o} W) and ~11^{o} in the meridional direction (6^{o} - 17^{o} N) and therefore a large fraction of the northern tropical Atlantic (Fig. 1). During August 2016, the intertropical convergence zone (ITCZ) was located in the southern part of the domain while the northern part mostly contains trade cumulus clouds. Hence, this case study provides an opportunity to study heterogenous clouds systems. Daily variations in the deep/shallow cloud modes in our domain were observed, but it always included both cloud modes, albeit in different relative fraction. Two different dates are chosen, one representing a shallow-cloud dominated mode (10-12/8/2016 – see Fig. 2, and Figs S1 and S3, supporting information- SI), and one that represents a deep-cloud dominated mode (16-18/8/16 – see Fig. 3 and Figs. S2 and S3, SI). In the shallow-cloud dominated case, most of the domain is covered by trade cumulus clouds that are being advected with the trade winds from north-east to south-west. In the southern part of the domain, throughout most of the simulation, there is a zonal band of deep convective clouds (Fig. 2) that contribute on average ~25% out of the total cloud cover (Fig. S3, SI). The deep-cloud dominated case represents the early stages of the development of the tropical storm Fiona (Fig. 3). Fiona formed in the eastern tropical Atlantic and moved toward the west-north-west. It started as a tropical depression at 16/8/2016 18:00 UTC while its centre was located at 12.0^{o} N 32.2^{o} W. It kept moving towards the north-west and reach a level of a tropical storm at 17/8/2016 12UTC, while its centre was located at 13.7^{o} N 36.0^{o} W (https://www.nhc.noaa.gov/data/tcr/AL062016_Fiona.pdf). The general propagation speed and direction, strength (measure by maximal surface wind speed) and location of the storm are predicted well by the model. However, the model produces more anvil clouds than what was observed from the satellite (Fig. 3). These two different cases, representing different atmospheric energy budget initial state (see also Figs. 4 and 12 below), enable the investigation of the aerosol effect on the energy budget under different initial conditions.

We use a two-moment bulk microphysical scheme (Seifert and Beheng, 2006b). For each case, four different simulations with different prescribed cloud droplet number concentrations (CDNC) of 20, 100, 200, and 500 cm^{-3} are conducted. The different CDNC scenarios serve as a proxy for different aerosol ~~concentration~~ conditions (as the first order effect of increased aerosol concentration on clouds is to increase the CDNC, ~~(~~Andreae, 2009~~)~~)~~.~~. This also allows to separate the cloud response from~~and avoid~~ the uncertainties involved in the representation of the aerosols in numerical models (Ghan et al., 2011; Simpson et al., 2014; Rothenberg et al., 2018). However, it limits potential feedbacks between clouds and aerosols, such as the removal of aerosol levels by precipitation scavenging and potential aerosol effects thereon. In addition, the fixed CDNC framework does not capture the differences in aerosol activation ~~fraction~~ between shallow and deep clouds, due to differences in vertical velocity. Another aerosol effect that is not included in our simulations is the direct interaction between aerosol and radiation. In future work we plan to examine the mutual interaction between the microphysical effects and the direct aerosol radiative effects.

For calculation of the difference between high CDNC (polluted) conditions and low CDNC (clean) conditions, the simulations with CDNC of 200 and 20 cm^{-3} are chosen as they represent the range typically observed over the ocean (see for example the CDNC range presented in recent observational-based studies (Rosenfeld et al., 2019; Gryspeerdt et al., 2019)). Each simulation is conducted for 48 hours starting from 12 UTC. The horizontal resolution is set to 1200 m and 75 vertical levels are used. The temporal resolution is 12 sec and the output interval is 30 min. Interactive radiation is calculated every 12 min using the RRTM-G scheme (Clough et al., 2005; Iacono et al., 2008; Mlawer et al., 1997). We have added a coupling between the microphysics and the radiation to include the Twomey effect (Twomey, 1977). This was done by including the information of the cloud liquid droplet effective radius, calculated in the microphysical scheme, in the radiation calculations. No Twomey effect due to changes in the ice particles size distribution was considered due to the large uncertainty involved in the ice microphysics and morphology. Additional details, such as the surface and atmospheric physics parameterizations, are described in Klocke et al., (2017) and include an interactive surface flux scheme and ~~a~~ fixed sea surface temperature (SST). We note that using a fixed SST does not include feedbacks of aerosols on the SST evolution that could change the surface fluxes. However, due to the large heat capacity of the ocean, we do not expect the SST to dramatically change over the two days simulations.

For comparing the outgoing longwave flux from the simulations and observations we use imager data from the SEVIRI instrument onboard the Meteosat Second Generation (MSG) geostationary satellite (Aminou, 2002). The outgoing longwave flux is calculated using the Optimal Retrieval for Aerosol and Cloud (ORAC) algorithm (Sus et al. 2017; McGarragh, et al. 2017). Cloud optical (thickness, effective radius, water path) and thermal (cloud top temperature and pressure) properties are retrieved from ORAC using an optimal estimation-based approach. These retrievals and reanalysis profiles of temperature, humidity and ozone are then ingested into BUGSrad, a two-stream correlated-k broadband flux algorithm (Stephens et al., 2001) that outputs the fluxes at the top and bottom of the atmosphere and shown to have excellent agreement when applied to both active (CloudSat) and passive (Advanced Along Track Scanning Radiometer) satellite sensors compared to Clouds and the Earth's Radiant Energy System (Henderson et al. 2013; Stengel et al. 2019). In addition, off-line sensitivity radiative transfer tests using vertical profiles from our model were conducted with BUGSrad to identify the source of the differences in fluxes between clean and polluted conditions.

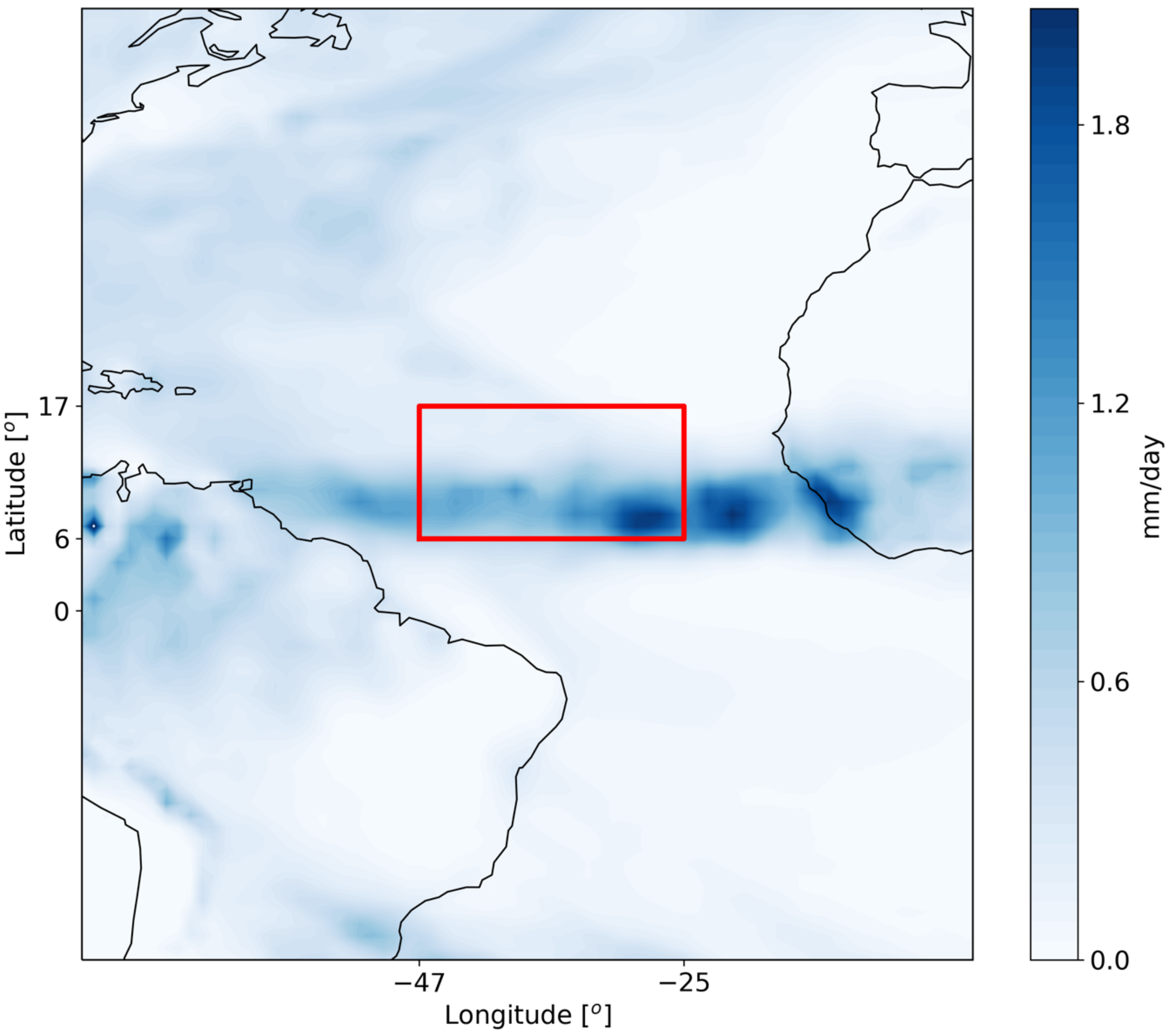


Figure 1. Domain of the ICON simulations (red rectangle) ~~for the NARVAL 2 case study~~ overlaid on the August 2016 ECMWF era-interim reanalysis (Dee et al., 2011) mean precipitation rate.

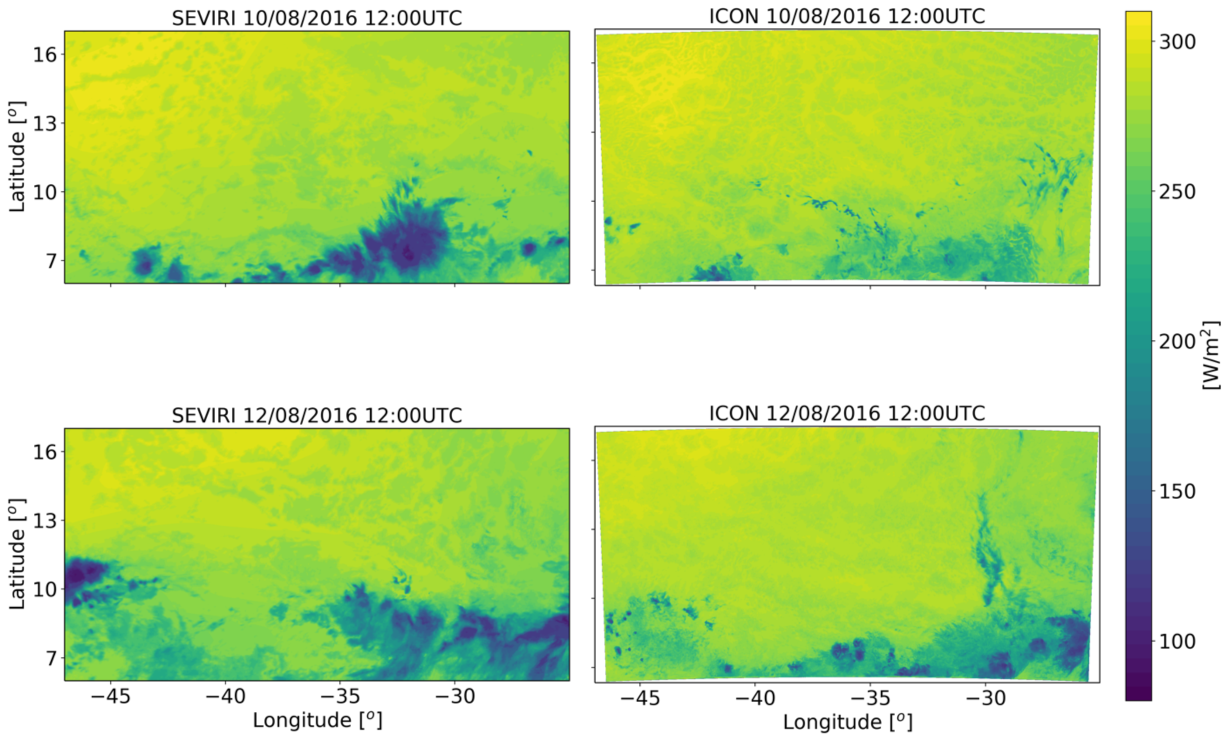


Figure 2. Outgoing longwave flux at the top of atmosphere at the initial stage (upper row) and the last stage (lower row – each average over 30 minutes) of the simulation of the shallow-cloud dominated case (10-12/08/2016) from geo-stationary satellite (SEVIRI-MSG – right column) and the ICON model simulation with CDNC of 20 cm^{-3} (left column).

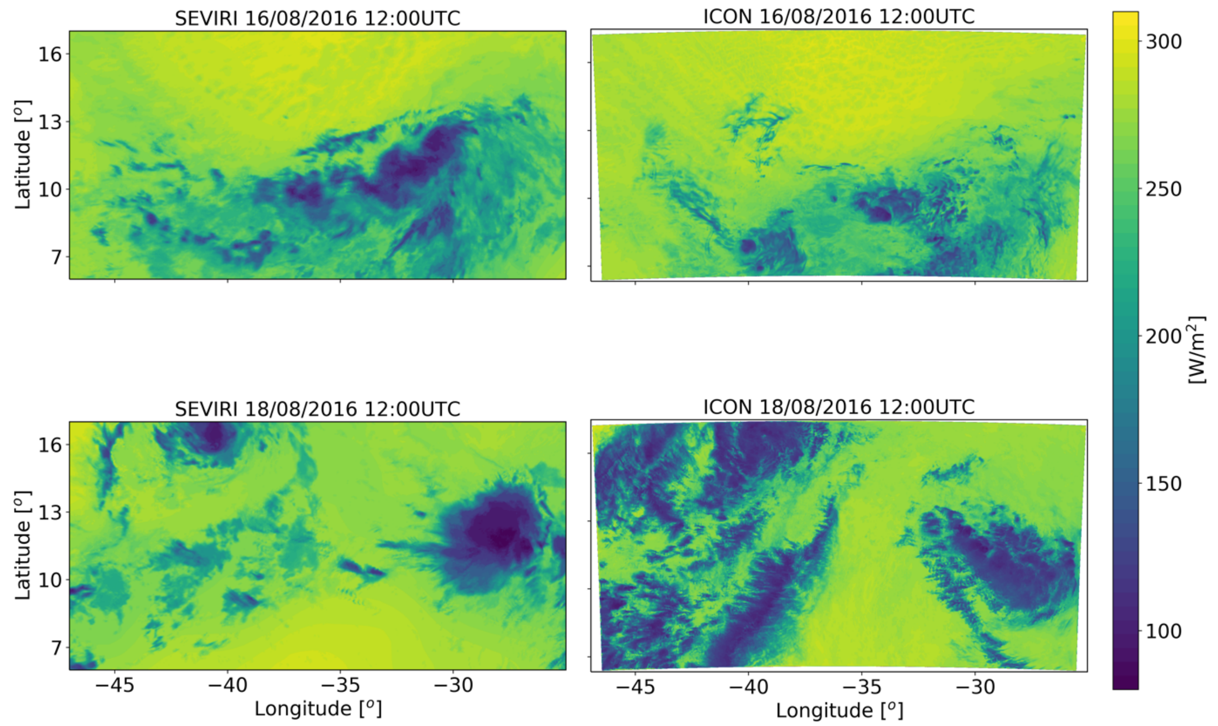


Figure 3. similar to Figure 2 but for the deep-cloud dominated case (16-18/08/2016).

Results

Shallow-cloud dominated case -10-12/08/2016

We start with energy budget analysis of the shallow-cloud dominated case base simulations (CDNC = 20 cm^{-3}). Figure 4 presents the time mean (over the two days simulation) of the different terms of the energy budget (Equation 1). As expected, *LP* dominates the warming of the atmosphere while Q_R dominate the cooling. The sensible heat flux (Q_{SH}) is positive (act to warm the atmosphere) but it is an order of magnitude smaller than the *LP* and Q_R magnitudes. In this shallow-cloud dominated case the radiative cooling of the atmosphere is significantly larger than the warming due to precipitation (mean of -114.7 W/m^2 compared to~~with~~ 90.1 W/m^2), hence the energy imbalance ~~residual~~ (R) is negative. Negative R means that there must be some convergence of dry static energy into the domain and/or decrease in the storage term, in this case it is mostly due to convergence of dry static energy.

We note that there is a significant difference in the spatial distribution of *LP* and Q_R (Jakob et al., 2019). While the Q_R is more uniformly distributed, the *LP* is mostly concentrated at the south part of the domain (where the deep convective clouds are formed) and it has a dotted structure. Locally, at the core of a deep convective clouds, the *LP* contribution can reach a few 1000 W/m^2 (1 mm/hr of precipitation is equivalent to 628 W/m^2), however, the vast majority of the domain contributes very little in terms of LP. Q_R also presents some spatial structure in which there is a weak atmospheric cooling at the south part of the domain (the region of the deep convective clouds) and a strong cooling at the reset of the domain.

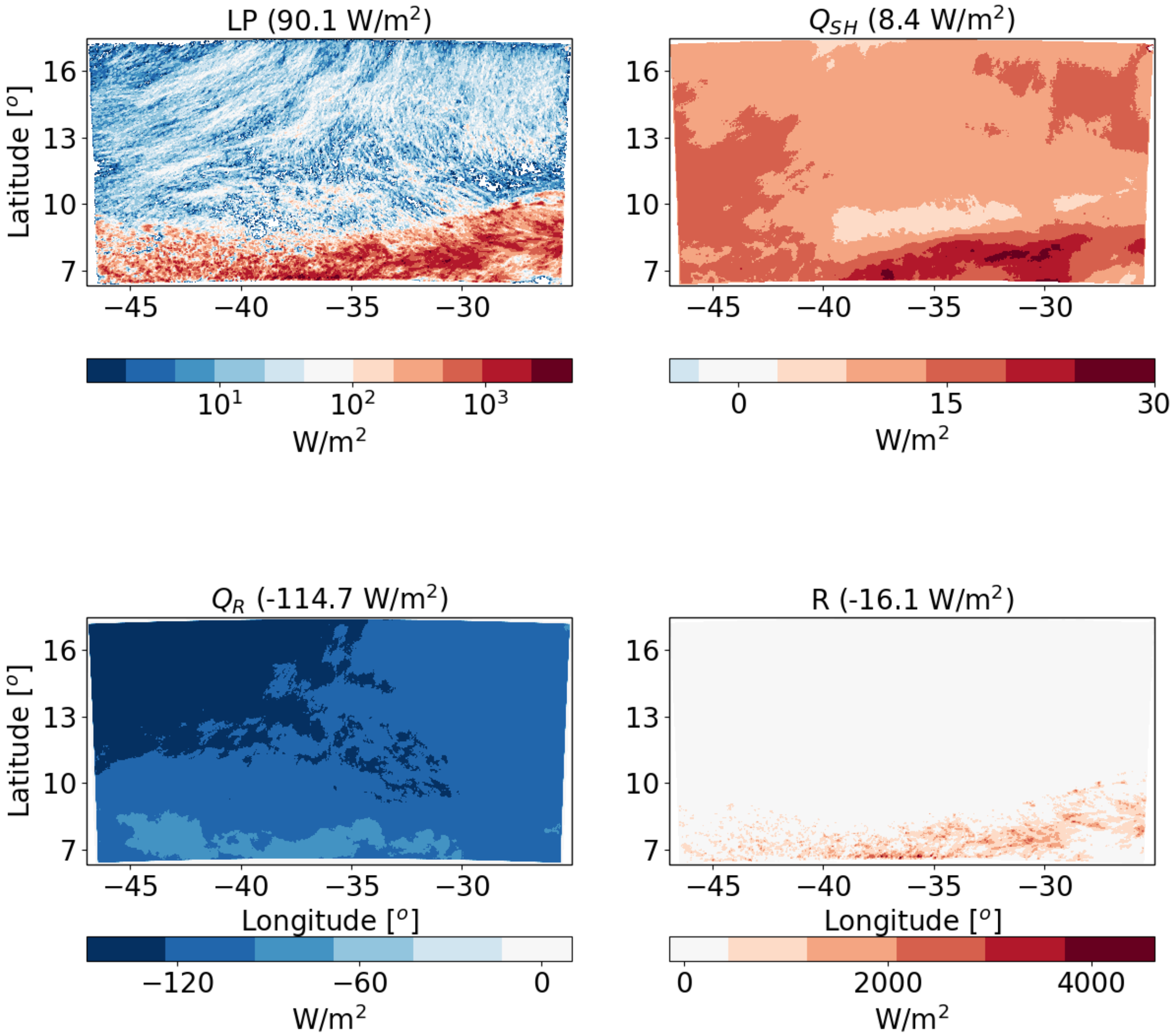


Figure 4. Spatial distribution of the time mean of the different terms of the energy budget for the ICON simulation of the shallow-cloud dominated case (10-12/08/2016) with CDNC = 20 cm^{-3}. The terms that appear here are: LP - latent heat by precipitation, Q_{SH} - sensible heat flux, Q_R - atmospheric radiative warming, and R – the ~~residual~~energy imbalance. The domain and time-mean value of each term appears in parenthesis.

For understanding the spatial structure of Q_R, next we examine the spatial distribution of the LW and SW radiative fluxes at the TOA and surface (Fig. 5). We note that the smaller radiative cooling in the region of deep clouds in the south of the domain is mostly contributed by a decrease in F_{LW}^{TOA}. The SW fluxes also demonstrate a strong south-north gradient, as the deep convective clouds in the south are more reflective than the shallow trade cumulus (with the lower mean cloud fraction) in the rest of the domain.

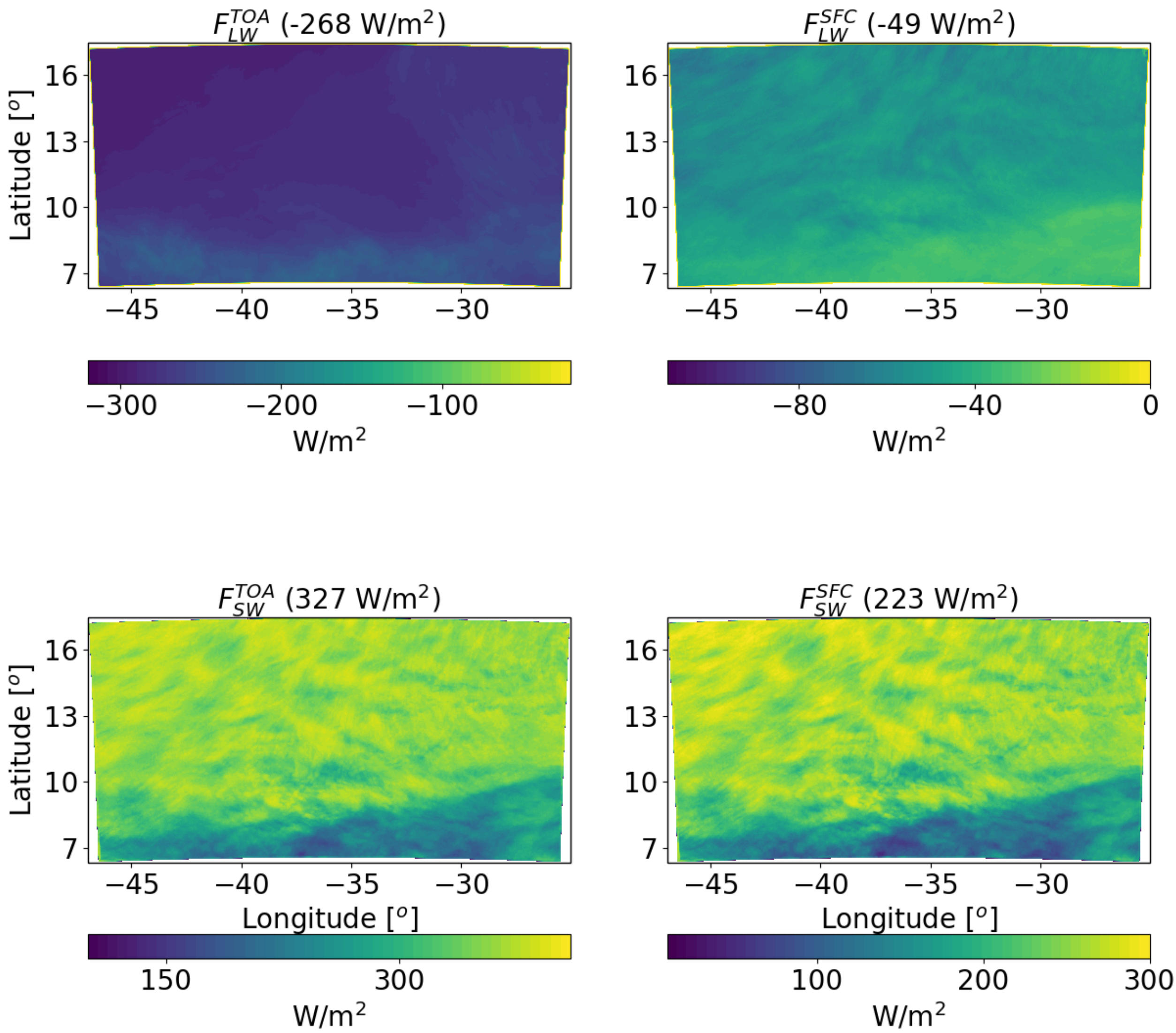


Figure 5. Spatial distribution of ICON simulated time-mean longwave (LW) and shortwave (SW) radiation fluxes at the top of atmosphere (TOA) and surface (SFC) for a simulation of the shallow-cloud dominated case (10-12/08/2016) with CDNC = 20 cm^{-3}. The domain and time mean value of each term appears in parenthesis.

<u>Response to aerosol perturbation – shallow-cloud dominated case</u>

Next, we analyse the response of the atmospheric energy budget of this case to perturbations in CDNC. Figure 6 presents the differences in the different terms of the energy budget between a polluted simulation (CDNC = 200 cm^{-3}) and a clean simulation (CDNC = 20 cm^{-3}). It demonstrates that the *LP* differences between the different CDNC scenarios contribute 5.1 W/m^2 less to warm the atmosphere in the polluted vs. the clean simulation. We note that this apparently large effect is caused by a small, non-statistically significant, precipitation difference (~0.4 mm over the two days of simulation - see Fig. 8 below). The strong sensitivity of the atmospheric energy budget to small precipitation changes (recalling that 1 mm/hr is equivalent to 628 W/m^2) exemplifies the caution one needs to take when looking on precipitation response in terms of energy budget perspective. The Q_R differences lead to relative warming of the atmosphere of the polluted case compared to the clean case by 1.6 W/m^2. We note that most of the Q_R differences are located in the south-west part of the domain. The Q_{SH} changes counteracts 1.4 W/m^2 of the atmospheric warming by Q_R and so the end result is a deficit of 4.8 W/m^2 in the atmospheric energy budget in the polluted simulation compared to the clean simulation. The decrease in the Q_{SH} is driven by an increase in the near surface air temperature (see Fig. 8).

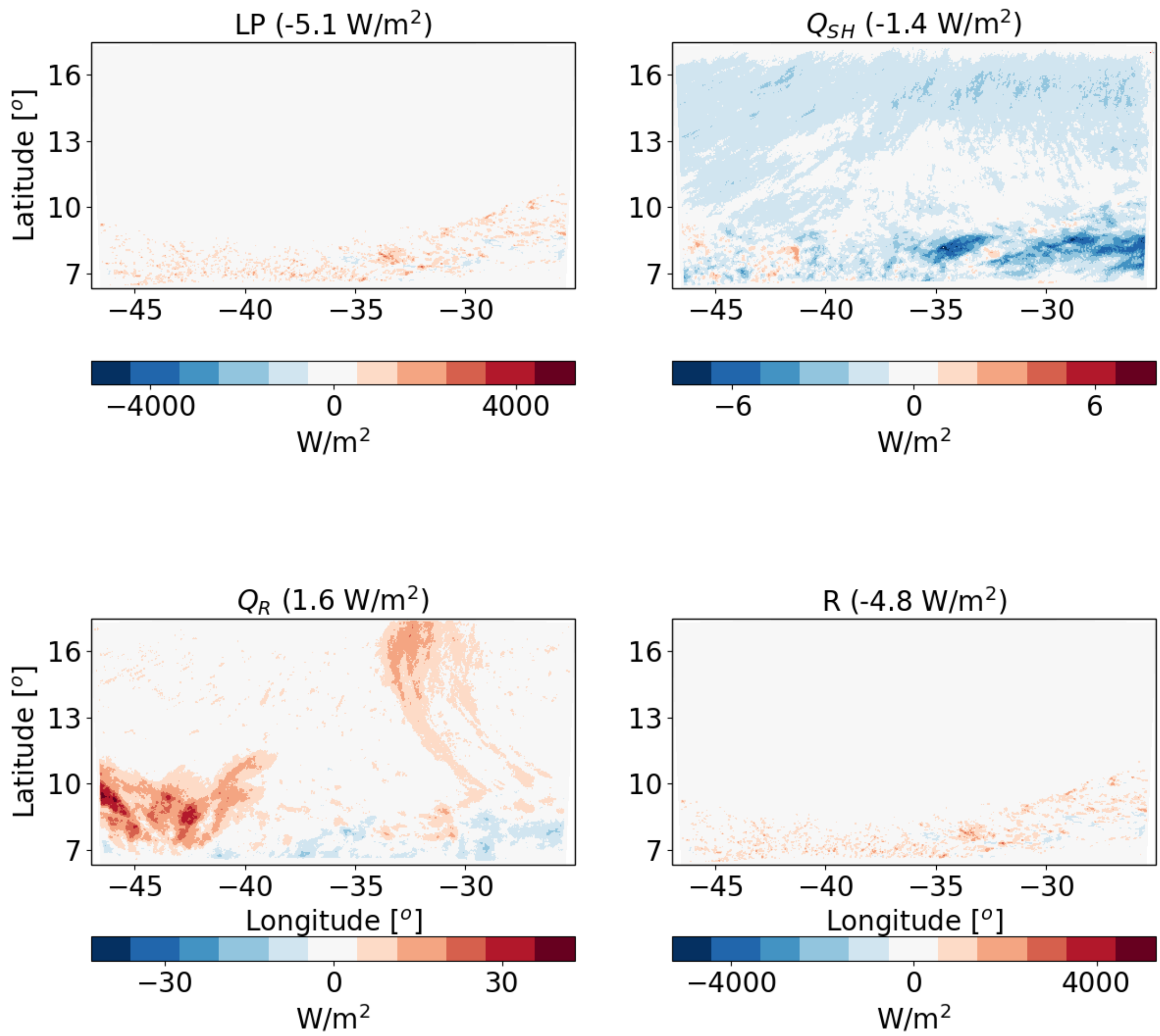


Figure 6. The differences between polluted (CDNC = 200 cm^{-3}) and clean (CDNC = 20 cm^{-3}) ICON simulations of the time-mean terms of the energy budget for the shallow-cloud dominated case (10-12/08/2016). The terms that appears here are: *LP* - latent heat by precipitation, Q_{SH} - sensible heat flux, Q_R - atmospheric radiative warming, and R – the ~~residual~~energy imbalance. The domain and time mean value of each term appears in parenthesis.

To understand the response of Q_R to the CDNC perturbation, we next examine the response of the different radiative fluxes. Figure 7 demonstrates that most of the relative atmospheric radiative heating in the polluted case compared to the clean case is contributed by changes in the F_{LW}^{TOA} fluxes. The changes in F_{LW}^{SFC} are an order of magnitude smaller. The SW fluxes change both at the TOA and SFC are larger than the F_{LW}^{TOA}changes, however, in terms of the atmospheric energy budget, they almost cancel each other out and the net SW atmospheric effect is only -0.9 W/m^2. Most of the reduction in SW fluxes (both at TOA and the surface) comes from the deep convective regions in the south of the domain while the shallow cloud regions experience some increase in SW fluxes. This can be attributed to the increase in deep convective cloud fraction and a decrease in the shallow cloud fraction with the increase in CDNC (see Fig. 9 below). The TOA net radiative effect for the entire system (as opposed to the atmospheric energy budget that take into consideration the surface radiative fluxes changes) is about -5.2 W/m^2.

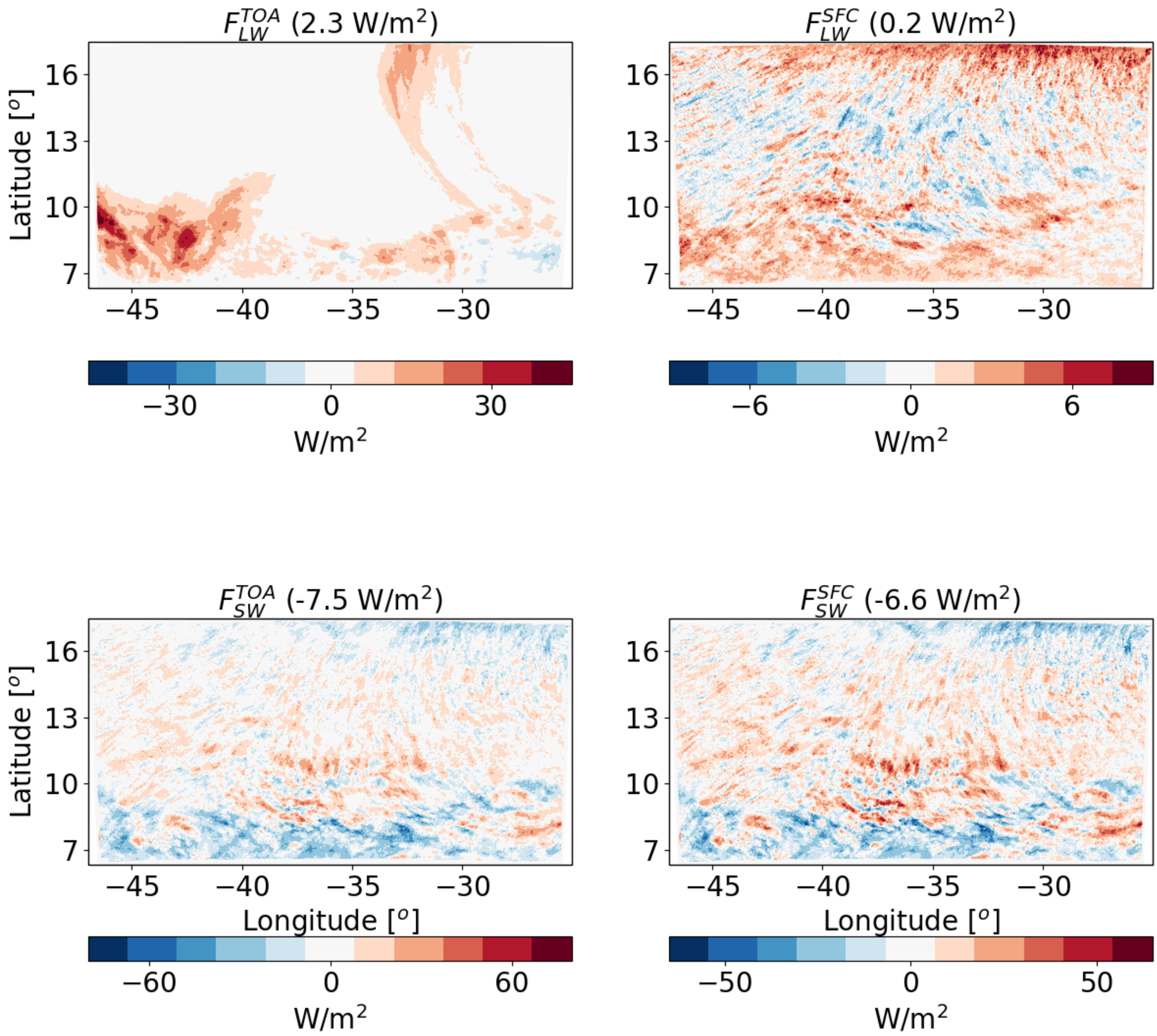


Figure 7. The differences between polluted (CDNC = 200 cm^{-3}) and clean (CDNC = 20 cm^{-3}) ICON simulations of the time mean radiative longwave (LW) and shortwave (SW) fluxes at the top of atmosphere (TOA) and surface (SFC) for the shallow-cloud dominated case (10-12/08/2016). The domain and time mean value of each term appears in parenthesis.

The differences in the energy (Fig. 6) and radiation (Fig. 7) budgets between the clean and polluted cases shown above, could be explained by the differences in the cloud mean properties. Figure 8 presents the time evolution of some of the domain mean properties while Fig. 9 presents time and horizontal mean vertical profiles. To examine the robustness of the trends we add here two more CDNC cases of 100 and 500 cm^{-3} (on top of the two that were examine above – 20 and 200 cm^{-3}). Figure 8 demonstrates that the domain mean cloud fraction (CF) generally decreases with the increase in CDNC (except for the first ~10 hours of the simulations). Examining the vertical structure of the CF response (Fig. 9), demonstrates that with the increase in CDNC there is a reduction in the low level (below 800 mb) CF concomitantly with an increase in CF at the middle and upper troposphere. The differences in rain rate between the different simulations are small. However, both the liquid water path (LWP) and the ice water path (IWP) show a consistent increase with CDNC. Accordingly, also the total water path (TWP), which is the sum of the LWP and the IWP, substantial increases with CDNC. The vertical profiles of the different hydrometers (Fig. 9) indicate, as expected, that the cloud droplet mass mixing ration (qc - droplet with radius smaller than 40 μm) increases with CDNC, while the rain mass mixing ratio (qr - drops with radius larger than 40 μm) decreases due to the shift in the droplet size distribution to smaller sizes under larger CDNC conditions. As this case is dominated by shallow clouds, there exists only a comparably small amount of ice mixing ration (qi) (c.f. Fig. 17), but its concentration increases with the CDNC increase. The combined effect of the increase in CDNC is to monotonically increase the total water mixing ratio (qt) above 800 mb (Fig. 9). The relative increase in qt with CDNC becomes larger at higher levels.

The increase in cloud water with increasing CDNC can explain both the reductions in the net downward SW fluxes (both at TOA and surface) and the decrease in outgoing LW flux at TOA (Fig. 7), as it results in more SW reflection concomitantly with more LW trapping in the atmosphere (Koren et al., 2010). Another contributor to the SW flux reduction (more reflectance) at the TOA is the Twomey effect (Twomey, 1977), while, the decrease in the low-level CF compensates some of this effect. Here we present the outcome of these contradicting effects on the SW fluxes, which shows a reduction at both the TOA and surface (Fig. 7). For estimating the relative contribution of the Twomey effect compare to the cloud adjustments (CF and TWP effects) to the SW flux changes, we have re-run the simulations with the Twomey effect turned off (the radiation calculations do not consider the changes in effective radius between the different simulations). It demonstrates that without the Twomey effect the TOA SW difference is only -1.7 W/m^2 as compared to -7.5 W/m^2 with the Twomey effect, demonstrating the predominant role of the Twomey effect. For estimating the relative contribution of the changes in CF and ~~TWP~~ water content to the SW flux changes we have conducted off-line radiative transfer sensitivity tests. To quantify the water content~~TWP~~ radiative effect, we feed the same CF vertical profile from the model into the offline radiative transfer model BUGSrad, while allowing the water content~~TWP~~ vertical profile to change (and visa versa to compute the CF radiative effect). This approach demonstrates that the contribution from the small reduction in CF is negligible compared to the increased SW reflectance caused by the increased~~ in~~ water content~~TWP~~ (the effect of the reduction in CF compensate only about 1% of the effect of the increase in the water content).

We also note a monotonic increase in the near surface temperature with CDNC (see also Fig. 10 below). This trend can be explained by warm rain suppression with increasing CDNC that leads to less evaporative cooling (see the decrease in the total amount of water mass mixing ration just above the surface in Fig. 9, (Dagan et al., 2016; Albrecht, 1993; Seigel, 2014; Seifert and Heus, 2013; Lebo and Morrison, 2014)). In addition, it was shown that under polluted conditions the rain drops below cloud base are larger, hence evaporating less efficiently (Lebo and Morrison, 2014; Dagan et al., 2016). The increase in the near surface temperature drives the decrease in the Q_{SH} (Fig. 6).

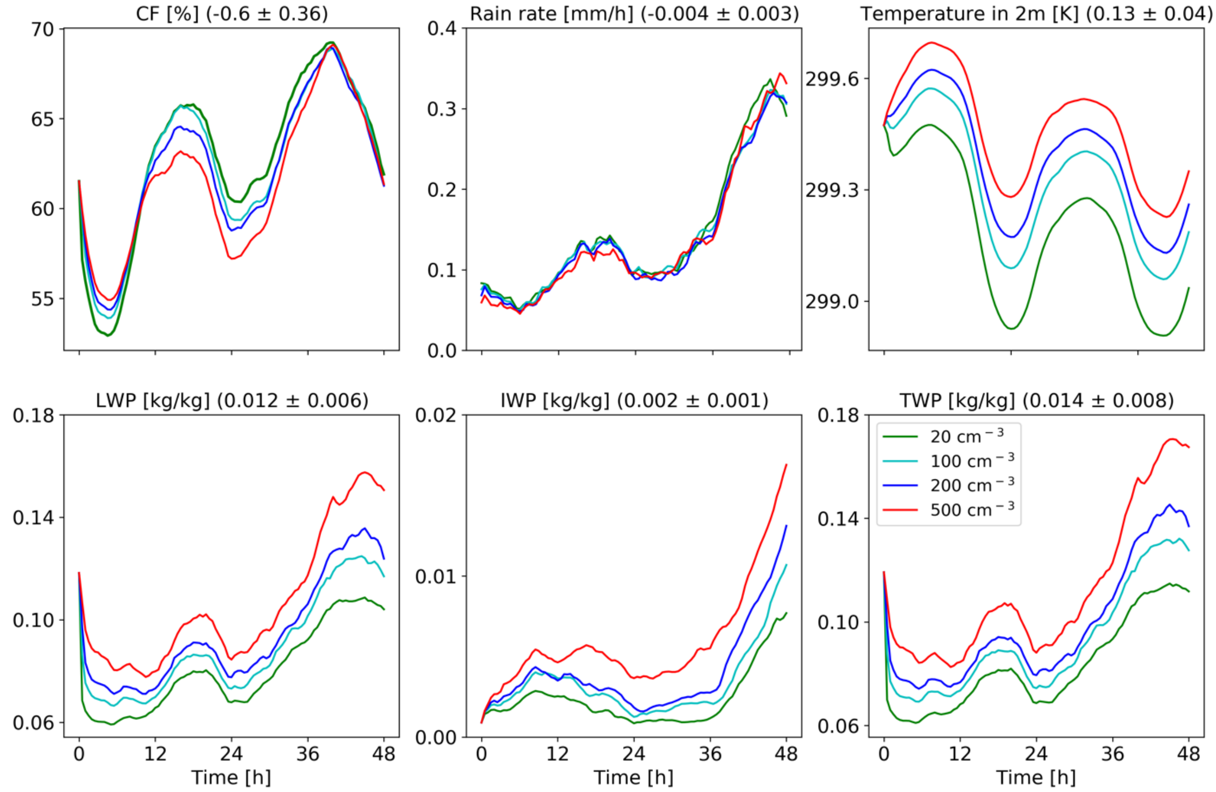


Figure 8. Domain average properties as a function of time for the different CDNC simulations for the shallow-cloud dominated case. The properties that are presented here are: cloud fraction (CF), rain rate, temperature in 2 m, liquid water path (LWP – based on the cloud water mass, excluding the rain mass for consistency with satellite observations), ice water path (IWP) and total water path (TPW = LWP + IWP). For each property, the mean difference between all combinations of simulations, normalized to a factor 5 increase in CDNC, and its standard deviation appear in parenthesis.

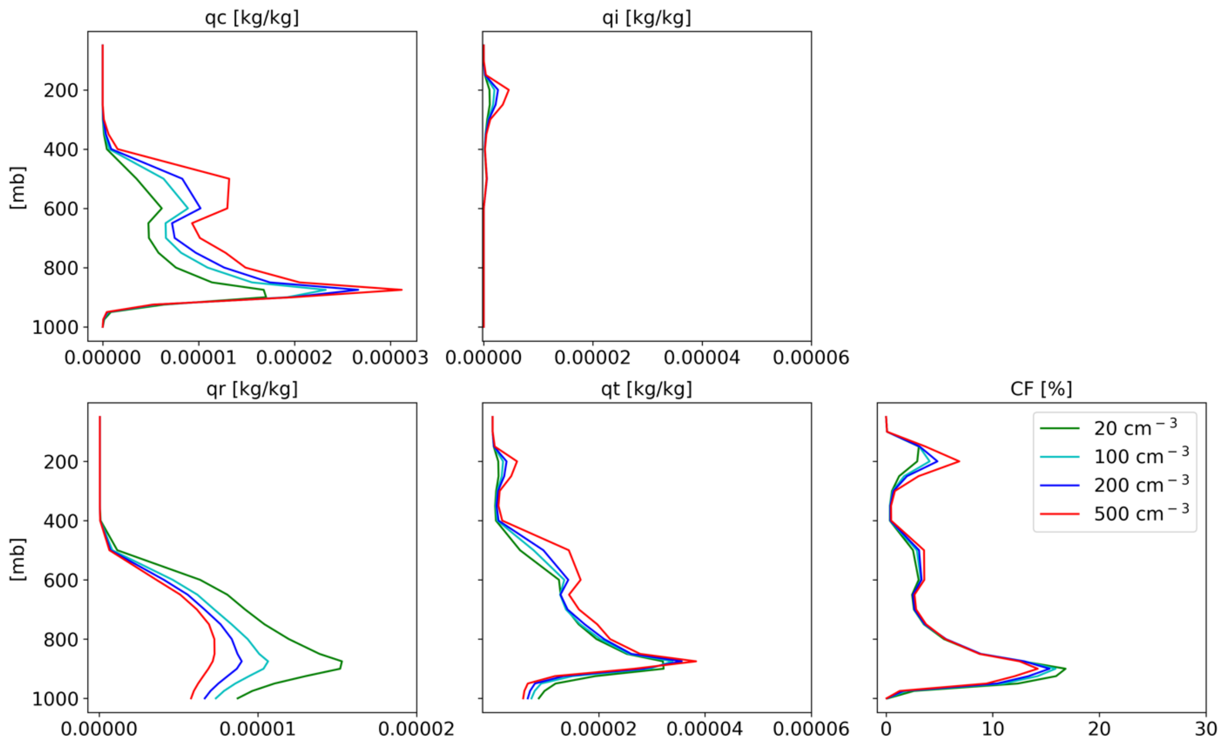
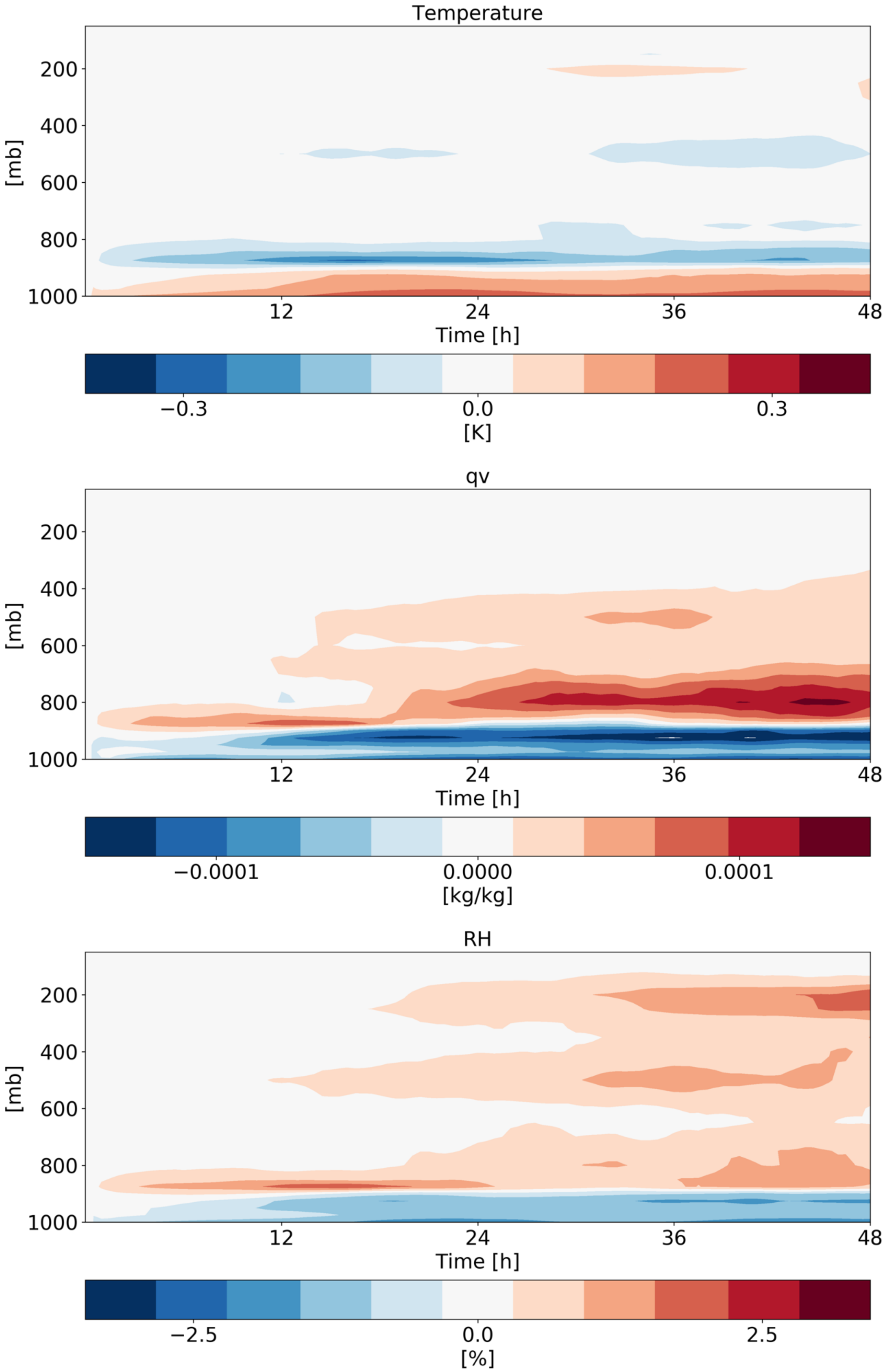


Figure 9. Domain and time average vertical profiles for the different CDNC simulations for the shallow-cloud dominated case. The properties that are presented here are: cloud droplet mass mixing ratio (qc – for clouds' droplets with radius smaller than 40 µm), ice mass mixing ratio (qi), rain mass mixing ratio (qr - for clouds' drops with radius larger than 40 µm), total water mass mixing ratio (qt = qc+qi+qr), and cloud fraction (CF). The x-axis ranges are identical as for the deep-cloud dominated case – Fig. 17.

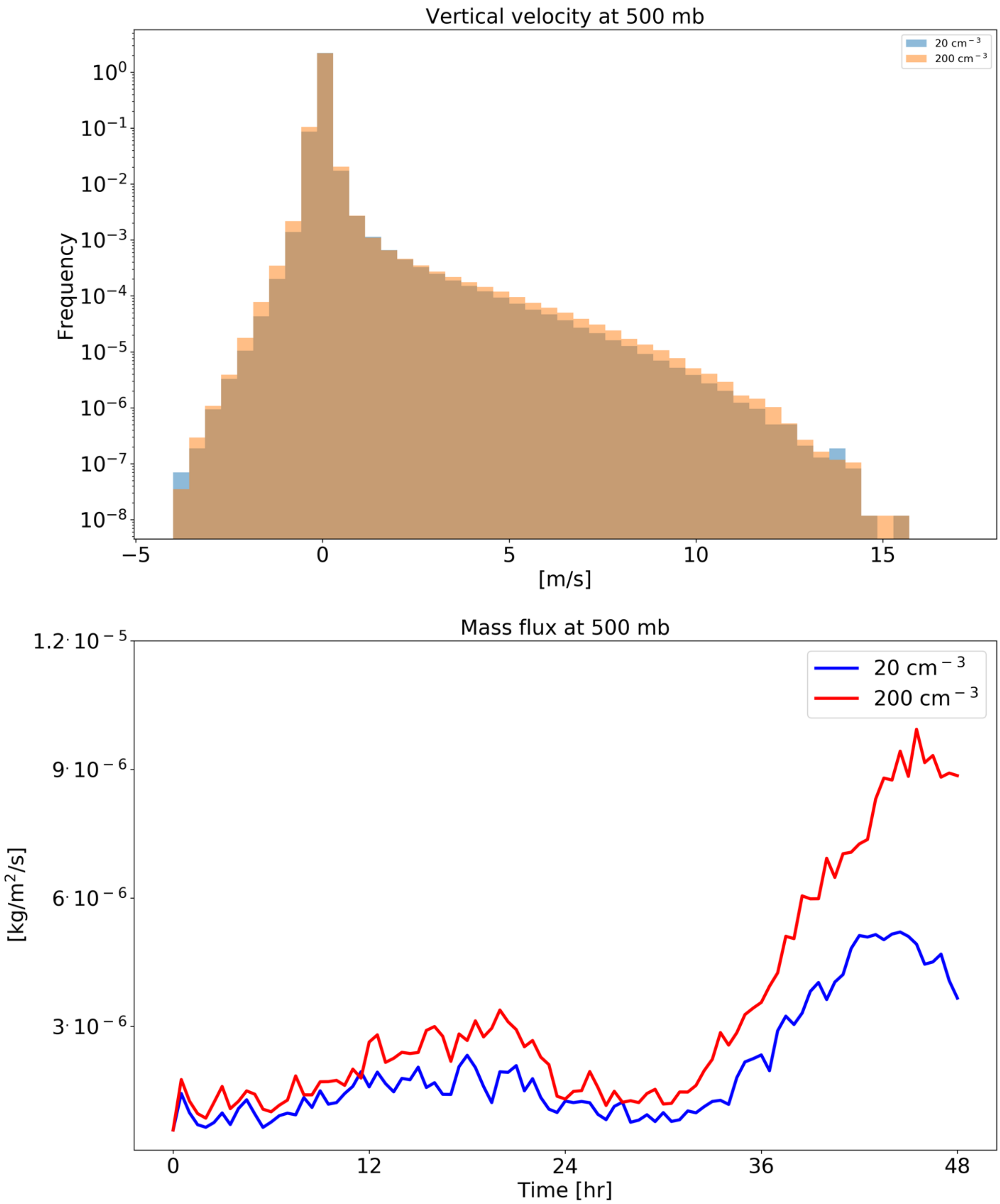
In addition to the clouds' effect on the radiation fluxes, changes in humidity could also contribute (Fig. 10). We note that increase in CDNC leads to increase in relative humidity (RH) and specific humidity (qv) at the middle and upper troposphere without a significant temperature change. The increased humidity at the upper troposphere would act to decrease the outgoing LW flux, ~~a~~ similar to the effect of~~as~~ the increased ice content in~~at~~ the upper troposphere~~ has~~ (Fig. 9). However, sensitivity studies with off-line radiative transfer calculations using BUGSrad~~,~~ demonstrate that the vast majority (more than 99%) of the different in F_{LW}^{TOA} between clean and polluted conditions emerges from the cloudy skies (rather than clear-sky), suggesting that the effect of the increased ice content at the upper troposphere dominates~~predominant~~.

Both the increase in water vapor and ice content in~~at~~ the upper troposphere are driven by an increase in upward water (liquid and ice) mass flux with increasing CDNC ~~to these levels~~ (Fig.

11). An increase in mass flux could be caused by an increase in vertical velocities and/or by an increase in cloud (or updraft) fraction and/or by an increase in cloud water content. In our case, t~~T~~he ~~increase~~increases in mass flux is driven partially by the small increase in vertical velocity (especially for updraft between 5 and 10 m/s – Fig. 11), partially by the small increase in cloud faction at this level (Fig. 9) and mostly due to the larger water mass mixing ratio (Fig. 9) that leads to an increase in mass flux even for a given vertical velocity. The increased relative humidity at the upper troposphere, further increases the ice particle lifetime at these levels (in addition to the microphysical effect (Grabowski and Morrison, 2016)) as the evaporation rate decreases. In addition, the differences in the thermodynamics evolution between the different simulations (Fig. 10) demonstrate drying and warming of the boundary layer with increasing CDNC, due to reduction in rain evaporation below cloud base and deepening of the boundary layer (Dagan et al., 2016; Lebo and Morrison, 2014; Seifert et al., 2015; Spill et al., 2019). The drying of the boundary layer could explain the reduction in the low cloud fraction (Fig. 9 (Seifert et al., 2015)).


Temperature
[mb]
200
400
600
800
1000
12
24
36
48
Time [h]
−0.3
0.0
0.3
[K]
qv
[mb]
200
400
600
800
1000
12
24
36
48
Time [h]
−0.0001
0.0000
0.0001
[kg/kg]
RH
[mb]
200
400
600
800
1000
12
24
36
48
Time [h]
−2.5
0.0
2.5
[%]

Figure~~-~~ 10. Time-height~~Hovmöller~~ **diagrams of the differences in the domain mean temperature, specific humidity (qv) and relative humidity (RH) vertical profiles between polluted (CDNC = 200 cm^{-3}) and clean (CDNC = 20 cm^{-3}) simulations for the shallow-cloud dominated case (10-12/08/2016).**


Vertical velocity at 500 mb
20 cm^{-3}
200 cm^{-3}
Frequency
10^{0}
10^{-1}
10^{-2}
10^{-3}
10^{-4}
10^{-5}
10^{-6}
10^{-7}
10^{-8}
−5
0
5
10
15
[m/s]
Mass flux at 500 mb
20 cm^{-3}
200 cm^{-3}
[kg/m^2/s]
$1.2 \cdot 10^{-5}$
$9 \cdot 10^{-6}$
$6 \cdot 10^{-6}$
$3 \cdot 10^{-6}$
0
12
24
36
48
Time [hr]

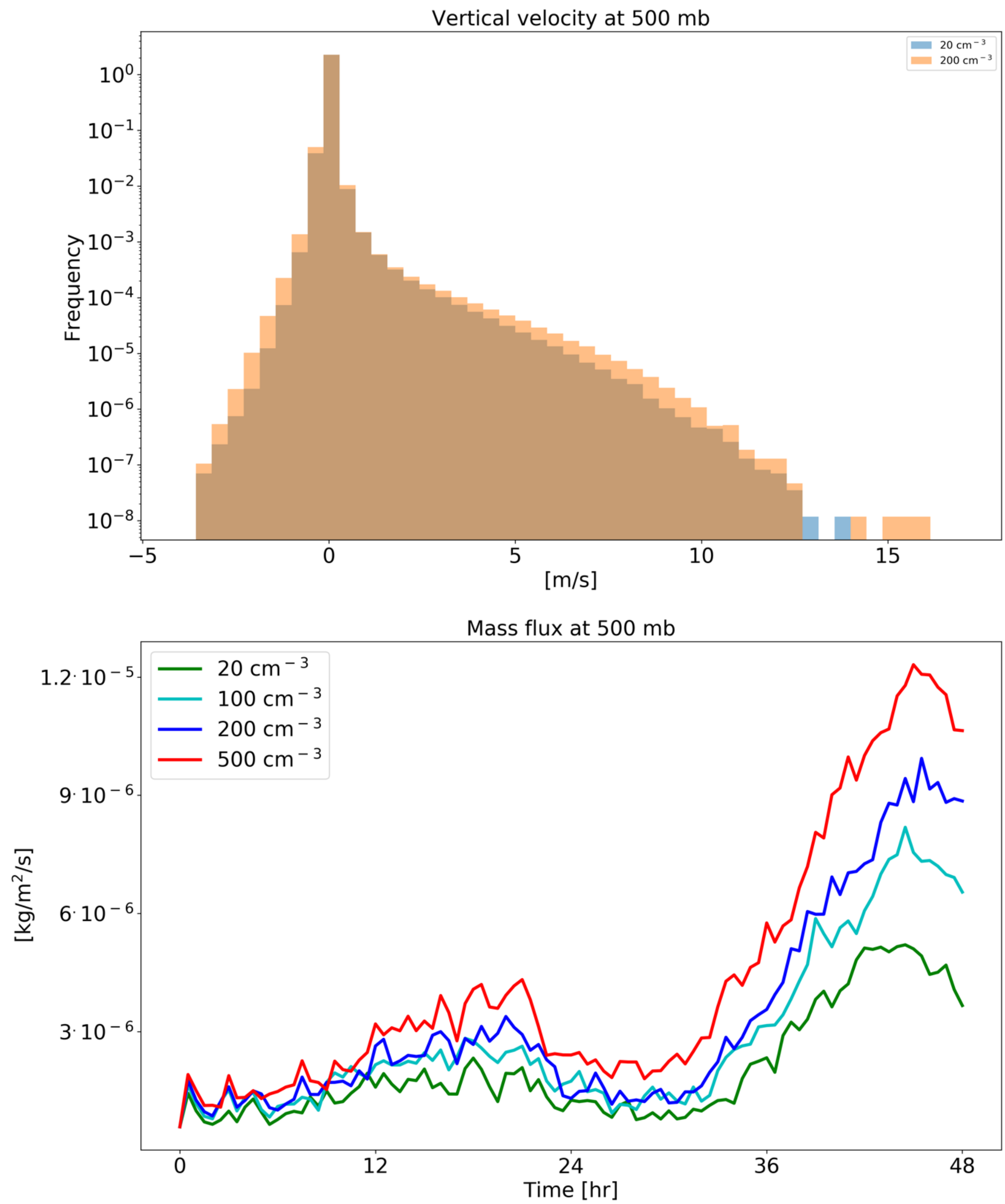


Figure 11. histograms of ICON simulated vertical velocity at the level of 500 mb for a clean (CDNC = 20 cm^{-3}) and polluted (CDNC = 200 cm^{-3}) simulations (upper), and the time evolution of the net upwards water (liquid and ice) mass flux (lower) for the different CDNC simulations ~~for a clean (CDNC = 20 cm^{-3}) and polluted (CDNC = 200 cm^{-3}) simulations~~ for the shallow-cloud dominated case (10-12/08/2016). The 500 mb level is chosen as it represents the transition between the warm part to the cold part of the clouds. In the histogram only two simulations are presented for clarity.

Deep-cloud dominated case -16-18/08/2016

Next, we analyse the atmospheric energy budget for the deep-cloud dominated case (Fiona tropical storm – Fig. 12). As opposed to the shallow-cloud dominated case, in this case the *LP* contribution dominates over the radiative cooling and hence the energy imbalance ~~residual~~ R is positive and large, suggesting divergence of dry static energy out of the domain. This difference in the base line atmospheric energy budget between the different cases simulated here, enable an examination of the aerosol effect on the atmospheric energy budget under contrasting initial conditions. As in the shallow-cloud dominated case, the Q_R values varies between small values (especially at the regions that were mostly covered by deep clouds) to larger negative values (dominated at the regions that were coved by shallow clouds). The Q_{SH} is positive and an order of magnitude smaller than the Q_R and *LP*, similar to the shallow-cloud dominated case.

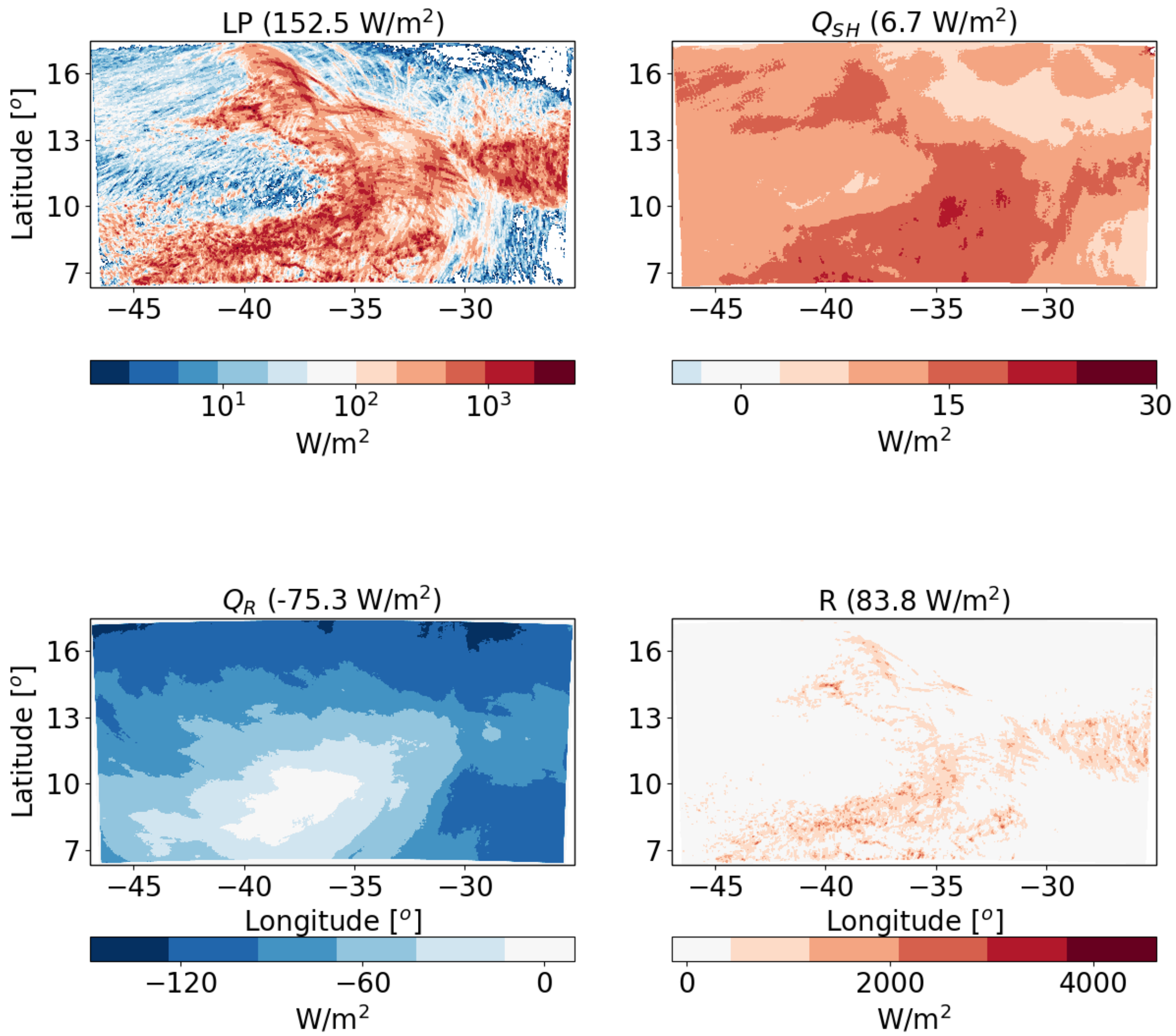


Figure 12. Spatial distribution of the time mean of the different terms of the energy budget for the ICON simulation of the deep-cloud dominated case (16-18/08/2016) with CDNC = 20 cm^{-3}. The terms that appear here are: *LP* - latent heat by precipitation, Q_{SH} - sensible heat flux, Q_R - atmospheric radiative warming, and R – the ~~residual~~energy imbalance. The domain and time-mean value of each term appears in parenthesis.

Further examination of the radiative fluxes (Fig. 13) demonstrates again the resemblance in the spatial structure between Q_R and $\boldsymbol{F_{LW}^{TOA}}$. As compared to the shallow-cloud dominated case, since the clouds are more opaque and cover larger fraction of the sky, there is a decrease in the magnitude of all fluxes (in different amount). For example, $\boldsymbol{F_{SW}^{SFC}}$ is lower by 41 W/m^2 (representing larger SW reflectance back to space) and the magnitude of $\boldsymbol{F_{LW}^{TOA}}$ by 47 W/m^2 as compare to the shallow-cloud dominated case. The combined effect of the radiative flux differences between the two cases is a decrease of the atmospheric radiative cooling by 39.6 W/m^2 (-114.7 compare with -75.3 W/m^2 – see Figs. 5 and 13).

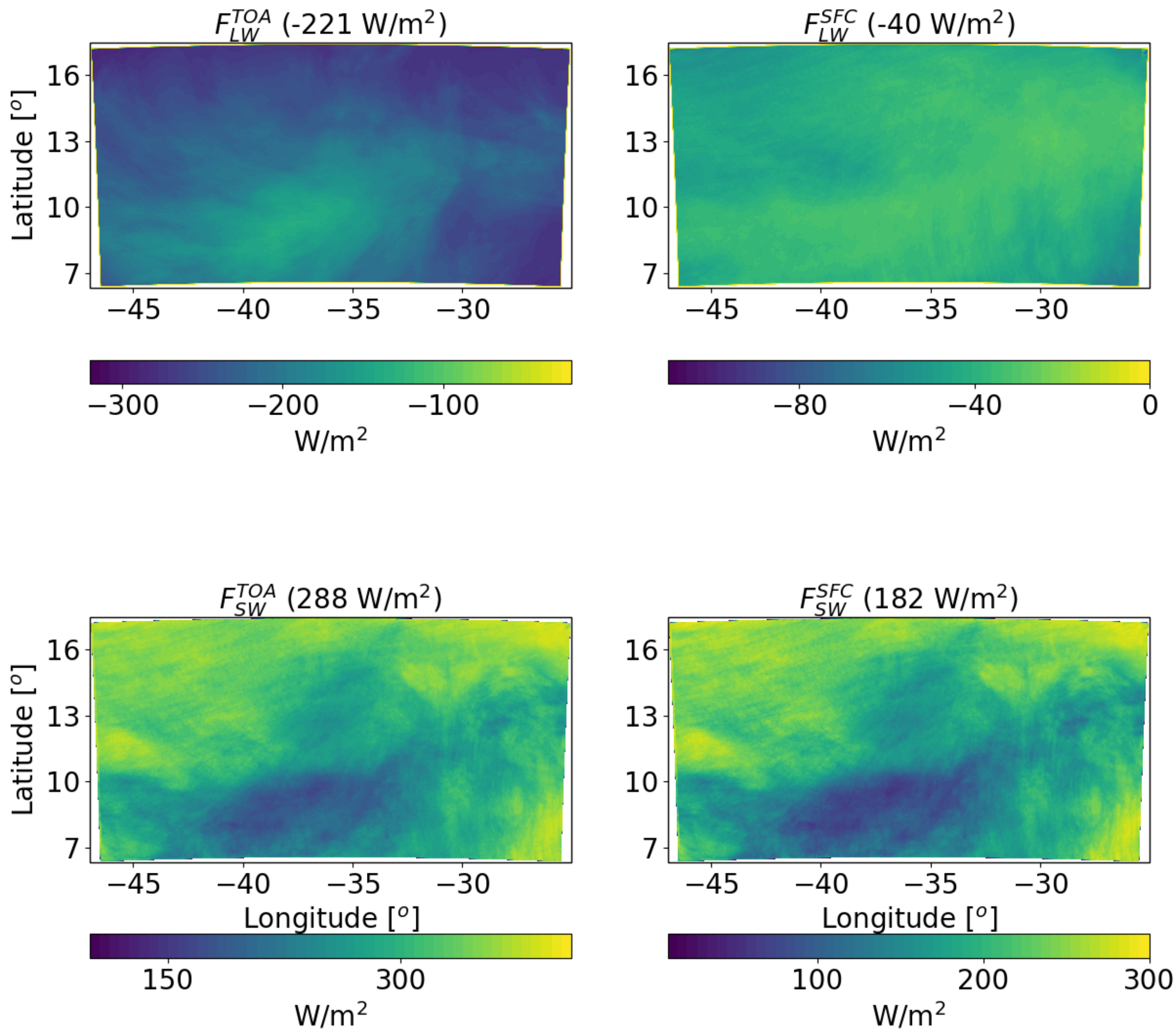


Figure 13. Spatial distribution of ICON simulated time-mean longwave (LW) and shortwave (SW) radiation fluxes at the top of atmosphere (TOA) and surface (SFC) for a simulation of the deep-cloud dominated case (16-18/08/2016) with CDNC = 20 cm^{-3}. The domain and time mean value of each term appears in parenthesis.

Response to aerosol perturbation – deep-cloud dominated case

For the deep-cloud dominated case, an increase in CDNC results in a decrease in *LP* by -0.3 W/m^2. Again, this difference is due to a non-statistically significant precipitation changes (see also Fig. 16 below). A similar Q_{SH} decrease as in the shallow-cloud dominated case is observed in the deep-clouds dominated case (see Figs. 14 and 6). The predominant difference in the response between the two cases is in Q_R, which increases much more in the deep-cloud dominated case: 10.0 W/m^2 (Fig. 14) compared with 1.6 W/m^2 in the shallow-cloud dominated case (Fig. 6).

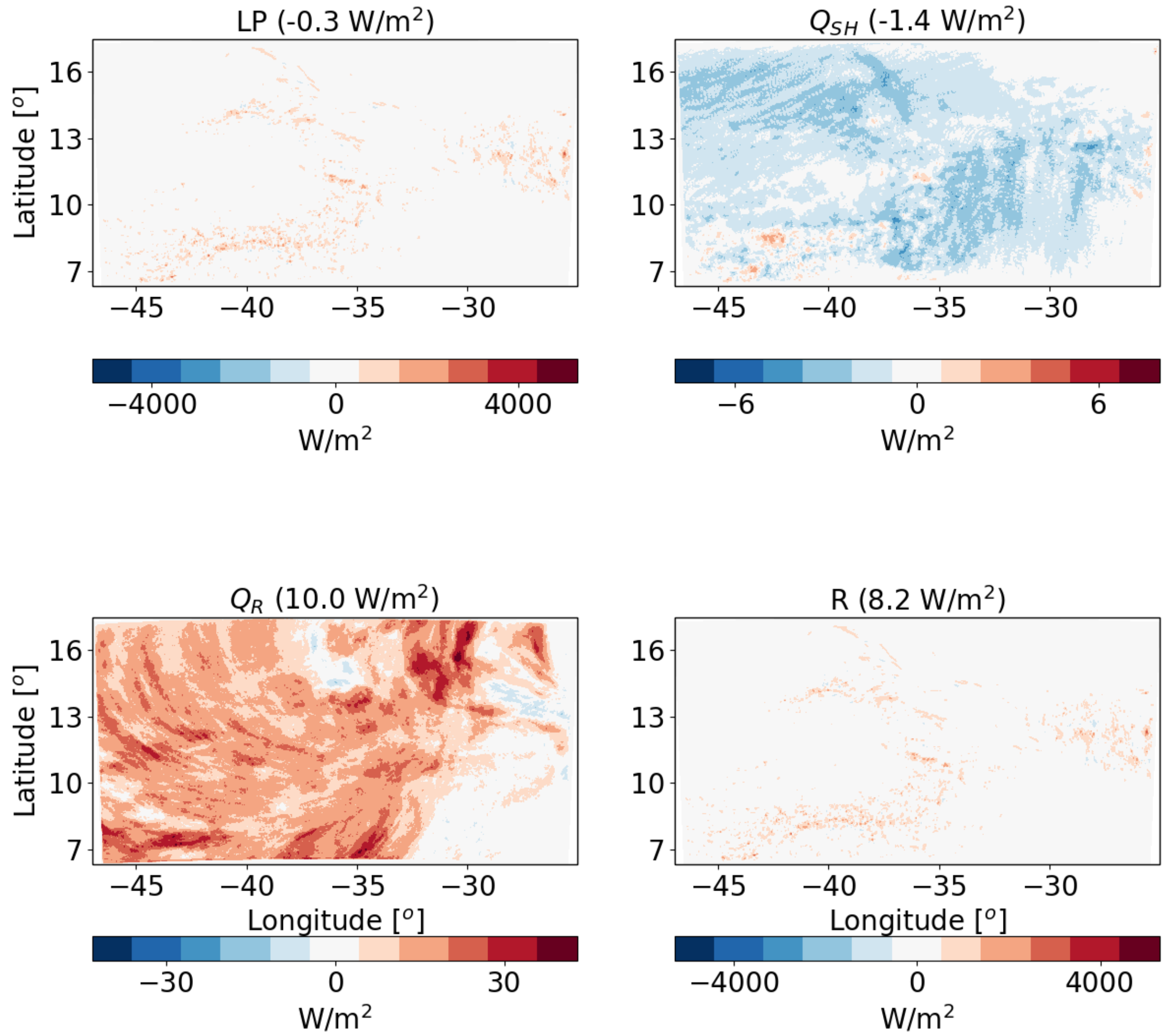


Figure 14. The differences between polluted (CDNC = 200 cm^{-3}) and clean (CDNC = 20 cm^{-3}) ICON simulations of the time-mean terms of the energy budget for the deep-cloud dominated case (16-18/08/2016). The terms that appears here are: *LP* - latent heat by precipitation, Q_{SH} - sensible heat flux, Q_R - atmospheric radiative warming, and R – the ~~residual~~energy imbalance. The domain and time mean value of each term appears in parenthesis.

The large increase in Q_R is caused mostly by the increase in F_{LW}^{TOA} (which becomes less negative i.e. less outgoing LW radiation under polluted conditions – Fig. 15). The CDNC effect on F_{LW}^{SFC} has a much smaller magnitude. The SW fluxes changes are substantial (-14.1 W/m^2 at TOA and -12.3 W/m^2 at the surface), however, in terms of the atmospheric energy budget, since clouds do not absorb much in the SW, the TOA and surface changes almost cancel each other out and the net effect is only ~1.8 W/m^2 atmospheric radiative cooling (which decrease some of the LW warming). The net TOA total (SW+LW) radiative flux change is about -1.9 W/m^2. The trends in the mean cloud properties (Figs. 16 and 17 below) can explain this large radiative response.

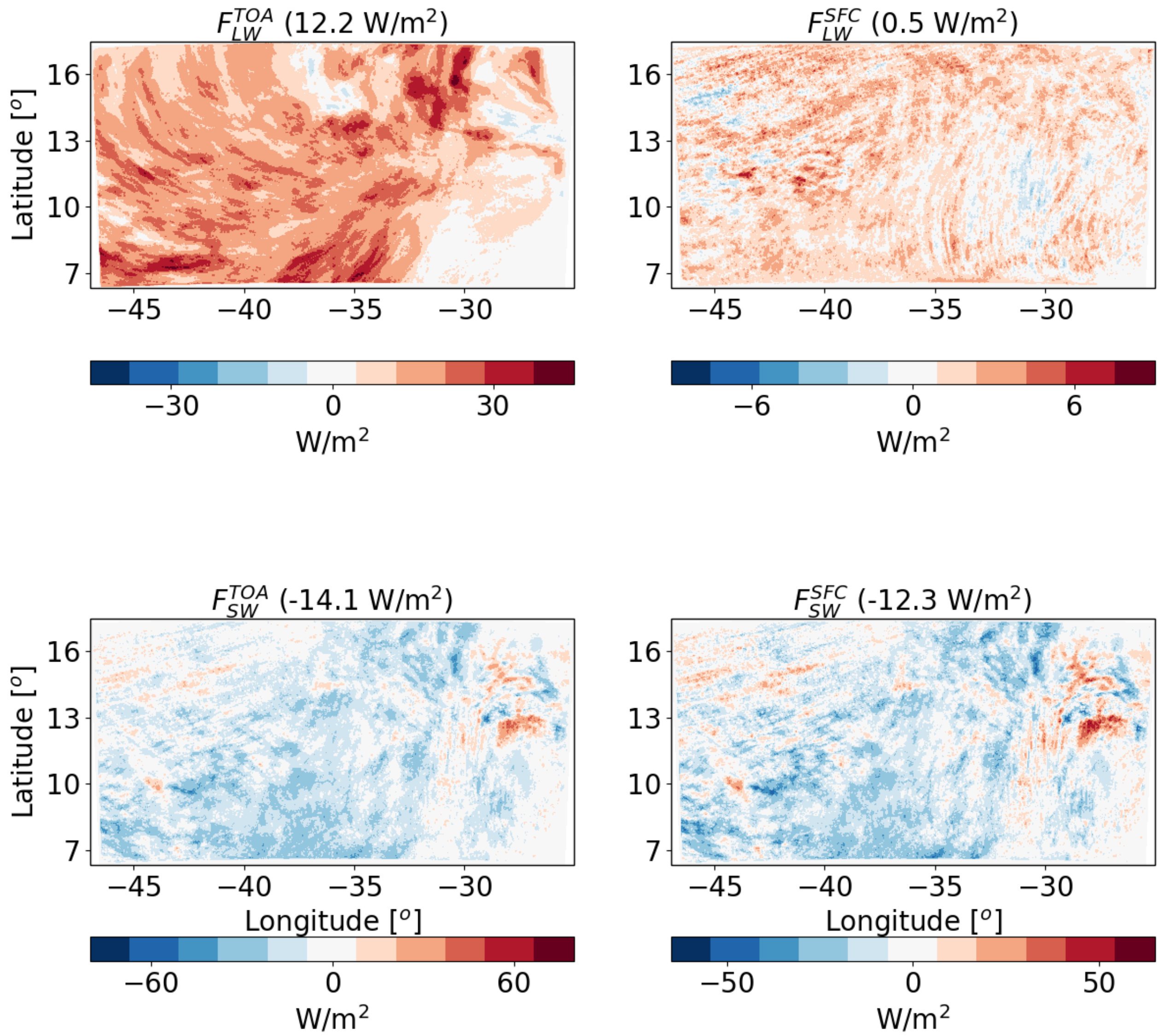


Figure 15. The differences between polluted (CDNC = 200 cm^{-3}) and clean (CDNC = 20 cm^{-3}) ICON simulations of the time mean radiative longwave (LW) and shortwave (SW) fluxes at the top of atmosphere (TOA) and surface (SFC) for the deep-cloud dominated case (16-18/08/2016). The domain and time mean value of each term appears in parenthesis.

Figure 16 presents some of the domain mean properties as a function of time for the deep-cloud dominated case. It demonstrates an increase in CF with CDNC which is more significant during the second day of the simulation. This is opposite to the CF reduction in the shallow-cloud dominated case (Fig. 8). It also demonstrates a very significant increase in LWP and, even more (in relative terms), in IWP and thus also in TWP. The increase in CF and water content can explain the decrease in SW fluxes both at TOA and surface (Fig. 15) as more SW is being reflected back to space. The larger SW reflection under increased CDNC is also contributed to by the Twomey effect (Twomey, 1977). Re-running the simulations without the Twomey effect result in 9.6 W/m^2 reduction in the TOA SW flux as compare to 14.1 W/m^2 with the Twomey effect on. We note that the relative role of the Twomey effect (compare to the cloud adjustments – CF and TWP) is larger in the shallow-cloud dominated case as compared to the deep-cloud dominated case (-14.1~~9.6~~ W/m^2 and -9.6~~14.1~~ W/m^2 for simulations with and without the Twomey effect in the deep-cloud dominated case, compare to -~~1.7~~7.5 W/m^2 and -1.7~~7.5~~ W/m^2 in the shallow-cloud dominated case, respectively). However, it should be noted that the Twomey effect due to changes in the ice particles size distribution was not considered. In this case, unlike in the shallow-cloud dominated case, the three contributions to the SW changes (CF, Twomey and LWP/IWP, e.g. (Goren and Rosenfeld, 2014)) all contribute to the SW flux reduction (Fig. 15 presents the results of all contributors). Off-line sensitivity tests demonstrate that the relative contribution of the ~~TWP~~ water content and the CF to the increase in SW reflectance is roughly ¾ and ¼, respectively.

The vertical profile changes with CDNC (Fig. 17) demonstrate a consistent picture of a decrease in CF in low clouds and a significant increase in CF and liquid and ice content at the mid and upper troposphere. The CF increase at the upper troposphere, and especially the increase in the ice content, can explain the decrease in the outgoing LW radiation (Fig. 15). The increase in ice content at the upper troposphere is in agreement with recent observational studies (Gryspeerdt et al., 2018; Sourdeval et al., 2018; Christensen et al., 2016). Analysis of the upward water mass flux from the warm to the cold part of the clouds (at 500 mb) in the different simulations (Fig. 19), demonstrates a substantial increase with the increase in CDNC (Chen et al., 2017), which occurs due to the increase in the water content (Fig. 17) and the delay in the rain formation to higher levels (Heikenfeld et al., 2019), even without a large change in the vertical velocity or cloud fraction at this level (Fig.17)~~due to the increase in the water content (Fig. 17) and the delay in the rain formation to higher levels (Heikenfeld et al., 2019)~~. Similar to the shallow-cloud dominated case (Fig. 8), the near surface temperature monotonically increases with CDNC, while the effect on the mean rain rate is small.

The differences in the thermodynamic evolution between polluted and clean conditions for this case (Fig. 18), demonstrate the same trend as in the shallow-cloud dominated case (Fig. 10). Here again, we note an increase in the humidity at the mid and upper troposphere, that contribute to the reduction in the outgoing LW flux. The deepening, drying and warming of the boundary layer are observed in this case as well. Both the increase in humidity at the mid-upper troposphere and the deepening of the boundary layer (Seifert et al., 2015) could cause a reduction of the outgoing LW flux. To distinguished the effect of clouds and humidity at the different levels on the outgoing LW flux, we have conducted sensitivity off-line radiative transfer calculations using BUGSrad. As in the shallow-cloud dominated case, the difference in outgoing LW flux between clean and polluted conditions primarily emerges from the CDNC effect on clouds. The small remaining effect of the clear sky (~0.2 W/m^2) is contributed by the change in the humidity at the mid and upper troposphere rather than by the deepening of the boundary layer (which would lead to LW emission from lower temperatures and is expected to be more significant under lower free troposphere humidity conditions).

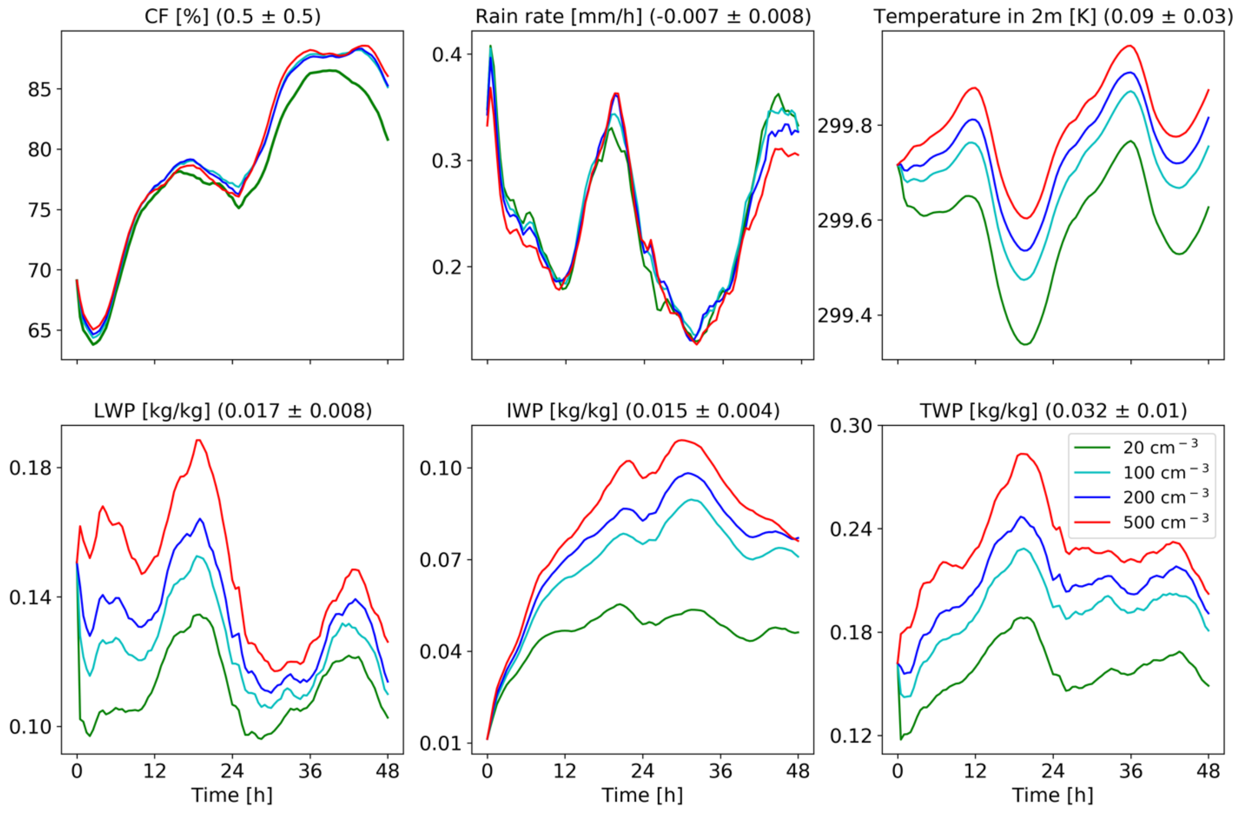


Figure 16. Domain average properties as a function of time for the different CDNC simulations for the deep-cloud dominated case. The properties that are presented here are: cloud fraction (CF), rain rate, temperature in 2 m, liquid water path (LWP – based on the cloud water mass, excluding the rain mass for consistency with satellite observations), ice water path (IWP) and total water path (TPW = LWP + IWP). For each property, the mean difference between all combinations of simulations, normalized to a factor 5 increase in CDNC, and its standard deviation appear in parenthesis.

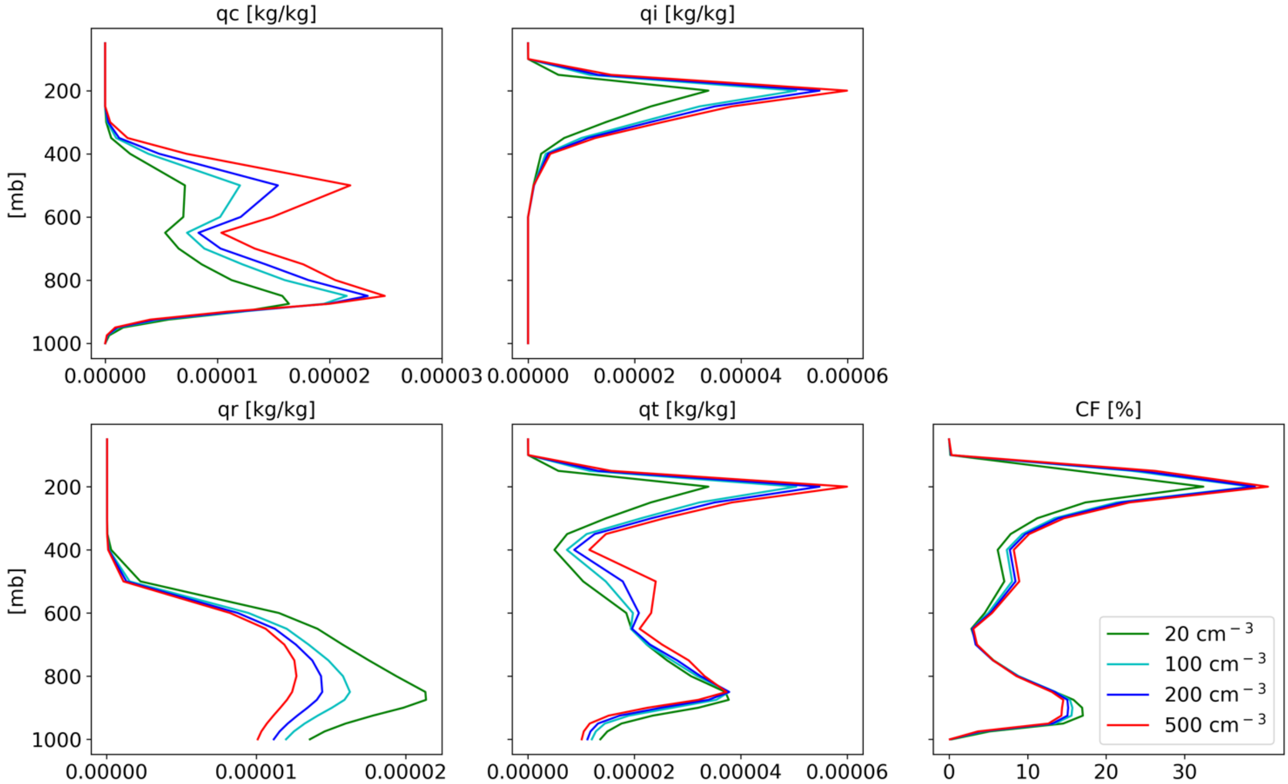


Figure 17. Domain and time average vertical profiles for the different CDNC simulations for the shallow-cloud dominated case. The properties that are presented here are: cloud droplet mass mixing ratio (qc – for clouds' droplets with radius smaller than 40 μm), ice mass mixing ratio (qi), rain mass mixing ratio (qr - for clouds' drops with radius larger than 40 μm), total water mass mixing ratio (qt = qc+qi+qr), and cloud fraction (CF).

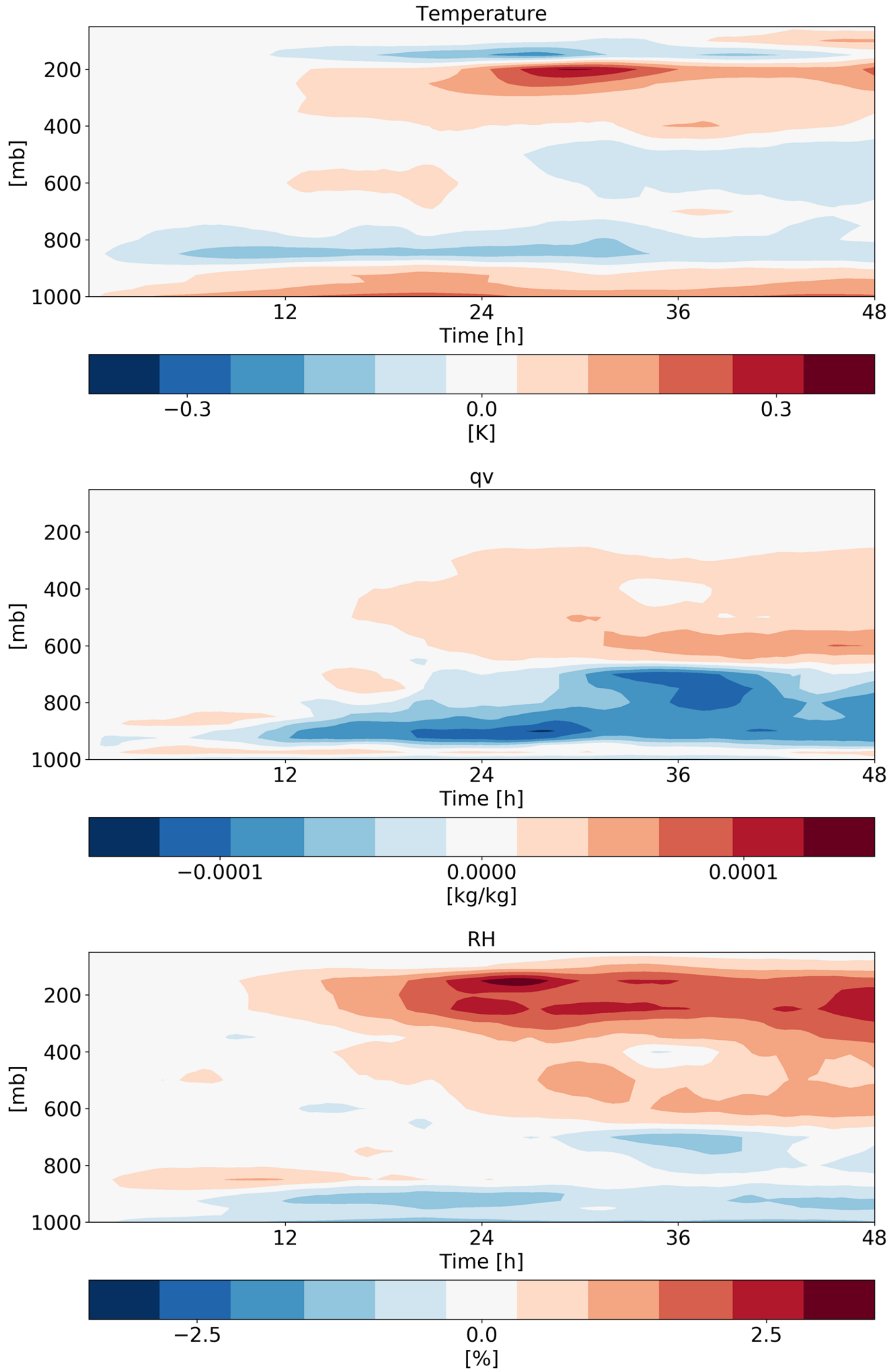
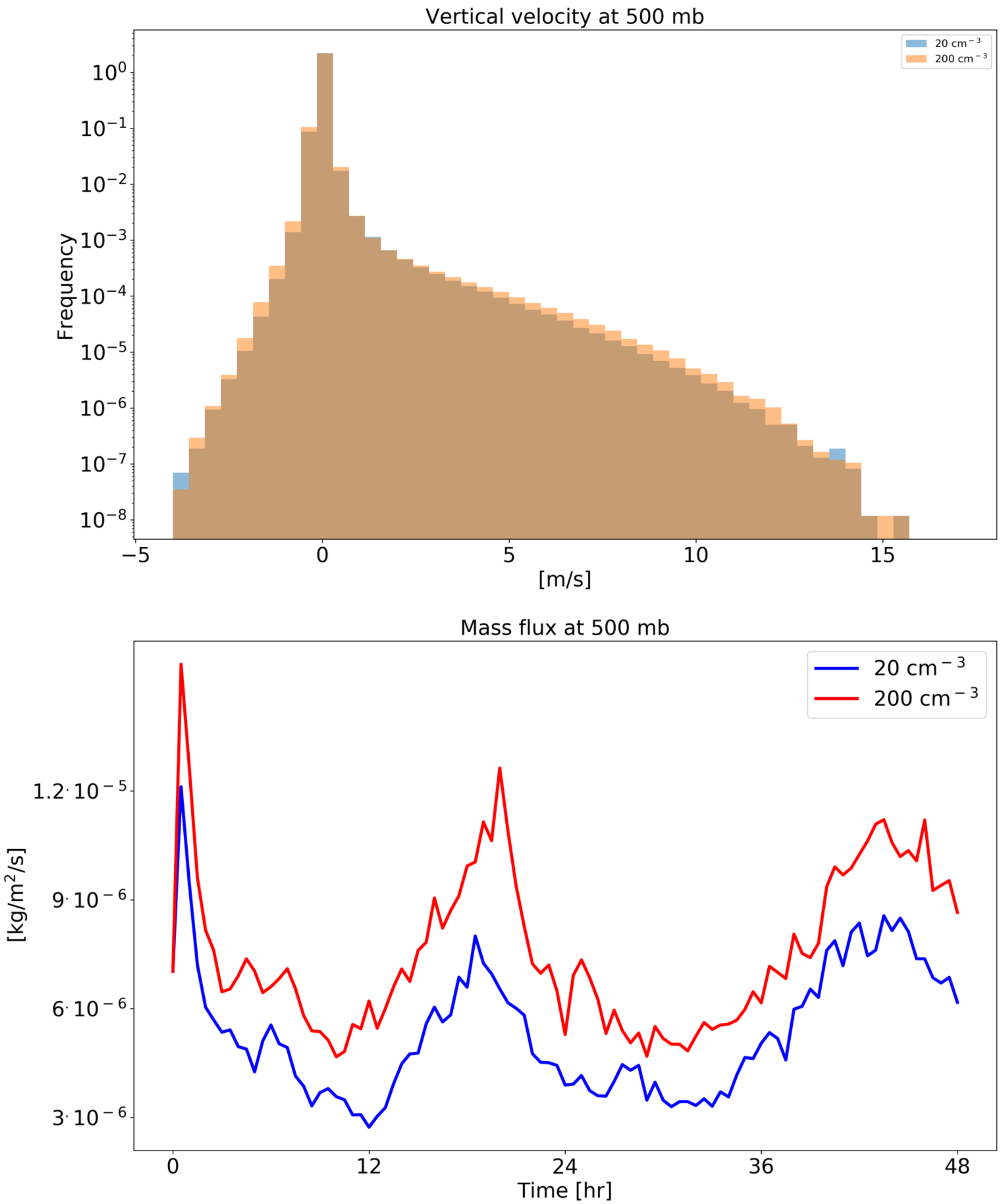


Figure 18. Time-height ~~Hovmöller~~ **diagrams of the differences in the domain mean temperature, specific**

humidity (qv) and relative humidity (RH) vertical profiles between polluted (CDNC = 200 cm^{-3}) and clean (CDNC = 20 cm^{-3}) simulations for the deep-cloud dominated case (16-18/08/2016).


Vertical velocity at 500 mb
Frequency
[m/s]
20 cm^{-3}
200 cm^{-3}
Mass flux at 500 mb
[kg/m^2/s]
Time [hr]
20 cm^{-3}
200 cm^{-3}

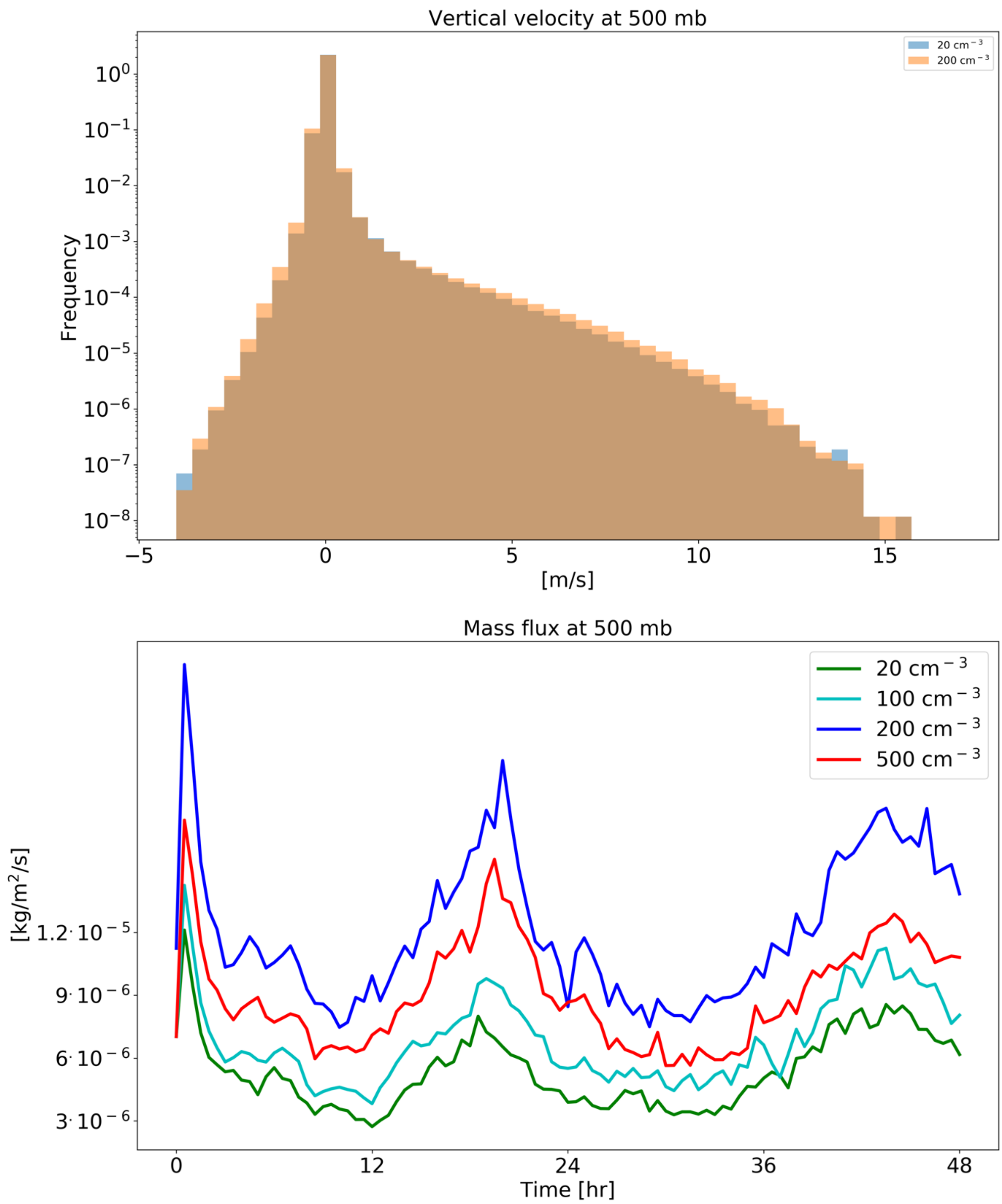


Figure 19.~~-~~ histograms of ICON simulated vertical velocity at the level of 500 mb for a clean (CDNC = 20 cm^{-3}) and polluted (CDNC = 200 cm^{-3}) simulations (upper), and the time evolution of the net upwards water (liquid and ice) mass flux (lower) for the different CDNC simulations for the deep-cloud dominated case (16-18/08/2016). The 500 mb level is chosen as it represents the transition between the warm part to the cold part of the clouds. In the histogram only two simulations are presented for clarity.~~histograms of ICON simulated vertical velocity at the level of 500 mb (upper), and the time evolution of the net upwards water (liquid and ice) mass flux (lower) for a clean (CDNC = 20 cm^{-3}) and polluted (CDNC = 200 cm^{-3}) simulations for the deep-~~

~~cloud dominated case (16-18/08/2016). The 500 mb level is chosen as it represents the transition between the warm part to the cold part of the clouds.~~

Summary and conclusions

Two different case studies of tropical cloud systems over the Atlantic Ocean were simulated using the ICON numerical model in a cloud resolving configuration with 1.2 km resolution and a relatively large domain (~22° x 11°). The cases represent dates from the NARVAL 2 field campaign that took place during August 2016 and have different dominant cloud types and different dominating terms in their energy budget. The first case (10-12/8/2016) is shallow-cloud dominated and hence dominated by radiative cooling, while the second case (16-18/8/2016) is dominated by deep convective clouds and hence dominated by precipitation warming. The main objective of this study is to analyse the response of the atmospheric energy budget to changes in cloud droplet number concentration (CDNC), which serve as a proxy for (or idealized representation of) changes in aerosol concentration. This enables better understanding of the processes acting in global-scale studies trying to constrain aerosol effect on precipitation changes using the energy budget perspective (O'Gorman et al., 2012; Muller and O'Gorman, 2011; Hodnebrog et al., 2016; Samset et al., 2016; Myhre et al., 2017; Liu et al., 2018; Richardson et al., 2018; Dagan et al., 2019a). Our results demonstrate that regional atmospheric energy budgets can be significantly perturbed by changes in CDNC and that the magnitude of the effect is cloud regime dependent (even for a given geographical region and given time of the year as the two cases are separated by less than a week).

Figure 20 summarizes the energy and radiation response of the two simulated cases to CDNC perturbations. It shows that the atmosphere in the deep-cloud dominated case experiences a very strong atmospheric warming due to an increase in CDNC (10.0 W/m^2). Most of this warming is caused by a reduction in the outgoing LW radiation at the TOA. The SW radiative fluxes (both at the TOA and surface) is also significantly modified but their net effect on the atmospheric column energy budget is small. The net TOA radiative fluxes change in this case is -1.9 W/m^2. Beside the atmospheric radiative warming, changes in precipitation (~-0.3 W/m^2), and in sensible heat flux (Q_{SH}, -1.4 W/m^2) also contribute to the total trend as a response of increase in CDNC. We note that since 1 mm/hr of rain is equivalent to 628 W/m^2, even negligible changes in precipitation of less than 0.5 mm over 48 hr (as seen in our simulations) can still appear as significant changes in the atmospheric energy budget and contribute a few W/m^2.

The response of the radiative fluxes can be explained by the changes in the mean cloud and thermodynamic properties in the domain. The mean cloud fraction (CF) increases with the increase in CDNC (Fig. 16) while the vertical structure of it indicates a reduction in the low cloud fraction (below 800 mb) and an increase in the mid and upper troposphere CF (Fig. 17). The water content (both liquid and ice) also increase with the increase in CDNC (Figs. 16 and 17) with increasing amount with height. These changes in the mean cloud properties drive both the reduction in SW fluxes at TOA and surface and LW flux at TOA as the clouds become more opaque (Koren et al., 2010; Storelvmo et al., 2011) and cover a larger fraction of the sky. In addition to cloud responses, the domain-mean thermodynamic conditions change as well (Fig. 18). Specifically, the humidity content at the mid and upper troposphere increases with higher CDNC, (due to increase mass flux to the upper troposphere) which further decreases the outgoing LW flux at the TOA. However, the vast majority of the LW effect emerges from the changes in clouds.

Both the increase in water vapor and ice content in~~at~~ the upper troposphere are driven by an increase in water mass flux with increasing CDNC to these levels (Fig. 19, (Koren et al., 2005; Rosenfeld et al., 2008; Altaratz et al., 2014; Chen et al., 2017)), which is caused mostly by the increase in the water mixing ratio in~~at~~ the mid-troposphere rather than by increase in vertical velocity (Figs.~~ 11 and~~ 19) or in cloud fraction (Fig. 17). The ice content in~~at~~ the upper troposphere is also increased due to reduction in the ice falling speed (Grabowski and Morrison, 2016), while the increased relative humidity at these levels, further increases the ice particle lifetime due to slower evaporation. However, the increase in water mass flux to the upper layers is not accompanied with an increase in precipitation as predicted by the classical “invigoration” paradigm (Altaratz et al., 2014; Rosenfeld et al., 2008), which suggest that some compensating mechanisms are operating (Stevens and Feingold, 2009).

In the shallow-cloud dominated case (which also contains a significant amount of deep convection), the response of Q_R is weaker but still substantial (a total decrease in the atmospheric radiative cooling of 1.6 W/m^2 - Fig. 20). The weaker total response under the shallow-cloud dominated conditions is due to the smaller role of the ice part in this case. Here again, the changes in Q_{SH} decrease about -1.4 W/m^2 of this atmospheric warming. As in the deep-cloud dominated case, most of the atmospheric radiative warming is caused by reduction in the outgoing LW flux, while the surface and TOA SW fluxes changes are non-negligible but cancel each other out (in terms of the atmospheric energy budget – reflecting small SW atmospheric absorption changes). However, a significant TOA net (SW+LW) radiative flux change of ~-5.2 W/m^2 remains. In this case, the cloud-mean effect on radiation is more complicated. While CF decreases with increasing CDNC, the mean water path (both LWP and IWP) increases (Fig. 8). As in the deep-cloud dominated case, the increase in the water content occurs mostly at the mid and upper troposphere, while the decrease in CF occurs mostly in the lower troposphere (Fig. 9). In terms of the SW fluxes, the effect of the decrease in low CF (decrease SW reflections) and the increase in water mass (increase SW reflections) would partially compensate, while the Twomey effect (Twomey, 1977) adds to the increase SW reflections. In this case, the net effect is more SW reflected back to space at TOA and a net negative flux change (including also the LW).

There exists a large spread in estimates of aerosol effects on clouds for different cloud types and different environmental conditions. In this study, as we use a relatively large domain (22^{o} x 11^{o}) and two different dates (each for two days), we sample many different local environmental conditions and cloud types. Such more realistic setups (although with lower spatial resolution) could provide more reliable estimates of aerosol effects on heterogeneous cloud systems than just one-cloud-type, small domain simulations (as was done in many previous studies, e.g (Dagan et al., 2017; Seifert et al., 2015; Ovchinnikov et al., 2014)). However, the conclusions demonstrated here are based on two specific cases. In order to examine the validity of our main conclusions over a wider range of initial conditions, we have conducted a large ensemble of simulations starting from realistic initial conditions (although with a smaller domain) in a companion paper (Dagan and Stier, 2019). These simulations demonstrate that the main conclusions presented in this paper are robust and hold also for a wide range of initial conditions representative for this area. In addition, the realistic setup with the continuously changing boundary conditions and systems that pass through the domain, which are used here, prevent conclusions that might be valid only in cyclic double periodic large eddy simulations, as the background meteorological conditions change more realistically (Dagan et al., 2018b). Another uncertainty in the assessment of the aerosol response are the large differences between different models and microphysical schemes (White et al., 2017; Fan et al., 2016; Khain et al., 2015; Heikenfeld et al., 2019). In this study, as we use only one model, we do not address this uncertainty. In future work we intend to examine the response in multiple models. In addition, more detailed observational constraints on the models are needed. Furthermore, we do not include the temporal evolution of the aerosol concentration. Feedbacks between the aerosol concentration and clouds processes (such as wet scavenging), as well as the direct effects of aerosol on radiation would add another layer of complexity that should be accounted for in future work.

Generally, the global mean aerosol radiative forcing is estimated to be negative (Boucher et al., 2013; Bellouin et al., 2019). However, these global aerosol forcing estimates have so far not included the radiative forcing associated with potential effects of aerosols on deep convection – and these effects are not represented in most current climate models due to limitations in convection parameterisations, with only a few exceptions (Kipling et al., 2017; Labbouz et al., 2018). Here we demonstrate the existence of non-negligible aerosol radiative effects (of -5.2 and -1.9 W/m^2 for the shallow and deep cloud dominated cases, respectively) in tropical cloud systems, that contained both deep and shallow convective clouds, with significant SW and LW contributions. From the (limited) two cases simulated here, it appears that (in agreement with previous studies) the aerosol effect may be regime dependent and that even within a given cloud regime the effect may vary with the meteorological conditions.

Finally, we hypothesise that the aerosol impact shown on the atmospheric energy balance, with increasing divergence of dry static energy from deep convective regions concomitantly with increased convergence in shallow clouds regions, can have effects on the large-scale circulation. This should be investigated in future work.

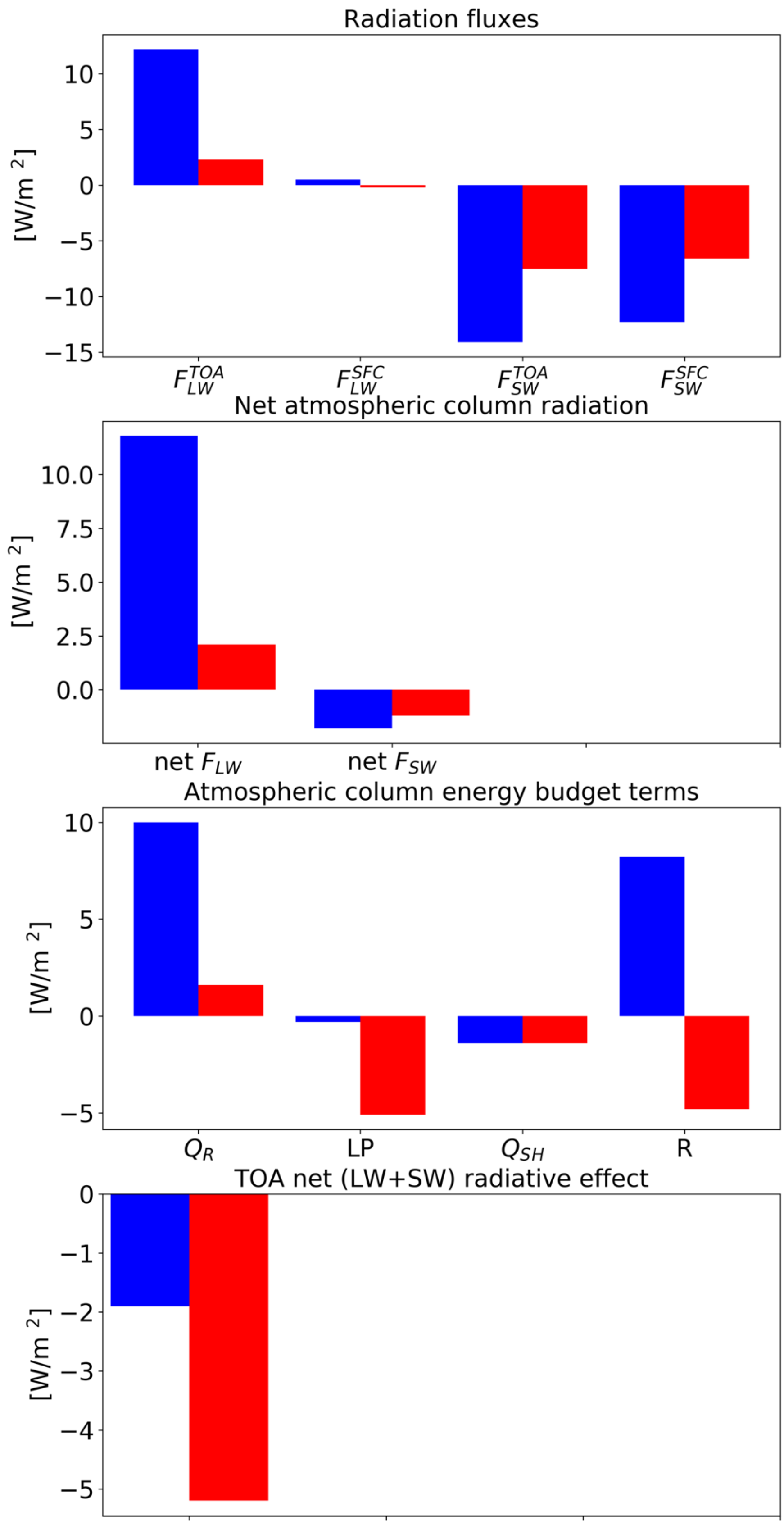


Figure 20. Summary of the radiation and energy response to CDNC perturbation in the two different cases. Blue represent the deep-cloud dominated case while red the shallow-cloud dominated case.

Author contributions. G. D. carried out the simulations and analyses presented. G.C., D.K. and A.S. assisted with the simulations. M.C. assisted with the radiative transfer calculations and comparison with observations. P. S. and A.S. assisted with the design and interpretation of the analyses. G. D. prepared the manuscript with contributions from all co-authors.

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