To the editor of ACP:

Herewith we submit the revised version of our manuscript. We are highly confident that we have addressed all comments of the two reviewers and that the revised manuscript will meet the high quality standards of ACP. The paper got improved in any respect. It got shorter by omitting material of secondary importance, and now includes deeper analysis.

We are looking forward to your response.

Best regards,

Ulrich Schumann and Bernhard Mayer, 22 August 2017

Responses to Reviewer 1.

We thank the reviewer for his thoughtful and detailed review. The review comments lead to considerable changes and several improvements.

This paper represents in some ways a rather impressive and stimulating study, but in its present form I am not sure its conclusions are safe ones (in the sense that they do not advance our understanding of real-world climate responses). I feel that the authors may be on to an important point, about the efficiency with which radiative perturbations in the upper troposphere can be transmitted to the surface, but whether the experiments presented here are sufficient to establish that importance is not so clear. I cannot give a strong recommendation for acceptance in anything like its present form. On the other hand I do not wish to discourage the authors from pursuing this important and interesting topic.

We have revised the paper in several parts. We tried to clarify open issues and to reduce the complexity by deleting any material which we feel is not absolutely necessary to bring over our central point: Mixing is important for climate sensitivity to contrail cirrus.

One of my issues with the paper is that it oscillates between being a fundamental study of the fate of radiative perturbations in the climate system and being a more applied and directed study concerning contrails in particular, and it is easy for the reader to get lost amongst material that is not clearly relevant. Certainly this reviewer felt lost on several occasions, and I found myself having to go back and re-read earlier sections and still I sometimes struggled. I am sorry to say that if I had not been a reviewer, I may not have persevered with reading the paper.

We now revised the paper at several places to reduce material and complexity and hope that the reviewer now finds it more worthwhile to spend time on the text.

So, for example, some of the approximations that are made may be appropriate to a more theoretical/illustrative study, are not so clearly appropriate if the aim is to specifically understand contrail efficacy. They might even invalidate the results. And similarly, while it might be useful to discuss the pure radiative equilibrium case in a theoretical study, that case is not really relevant to understanding contrail efficacy. I feel that the repeated presentation of the radiative equilibrium case gets in the way of understanding the real-world response. Overall, I felt the manuscript tried to be too "completist" (e.g. presenting figures and calculations that didn't need to be presented) which made the manuscript longer and more complex than it needed to be.

We agree in several respect and state now more clearly the merits and limitations of the approach, in Section 3 and in the Conclusions. On the other hand, we found that the results are robust against many model parameters, as now explained in subsections 3.1 and 3.2.

We deleted the discussion on the pure radiative equilibrium case and the related former figures 4 to 6.

A central issue in this paper is the ability of real-world cirrus/contrails to distort the vertical profile of temperature in the way that is shown in figures 5 and 6. It is this stabilization that is key to the authors' results. Is there any wide scale evidence that cirrus of contrails do this, particularly away from the rather special conditions in the tropical tropopause layer? I feel the authors need to do a critical analysis of the literature on this point, as the paper would be greatly strengthened if they are able to present any such evidence.

The stabilization is not the central issue (a lapse rate change is a classical feedback). The main point is the importance of mixing for surface temperature sensitivity, even for fixed lapse rate. Note, the figures on radiative-convective equilibrium were presented as test cases to demonstrate that the model passes necessary tests. We agree that convective adjustment is not so important for mid-latitude cirrus and reduced these parts considerably, therefore.

Detailed comments. Those preceded by an asterisk are more major comments.

1:1 I have a concern about the title. I do not believe that cirrus causes radiative forcing. It certainly has a radiative effect that can change (and hence induce a feedback) but this is rather unlike the contrail case. Perhaps "contrails and contrail cirrus" would be better as these are more obviously forcings.

We still mean that the study is of relevance beyond contrails, but now decided to change the title and to reduce the scope. So the new title refers more restrictively to contrail cirrus. In the Conclusions we still state: "The findings may apply also for other disturbances."

1:8 "basically without climate system changes" – presumably this means "no feedbacks" except for temperature change?

Yes, and the new wording says this.

1:13 "Heat induced by cirrus" – since in principle there is a latent heating associated with cirrus formation, clarify that this is "radiative heating due to cirrus"

We now talk about "energy induced by radiation".

1:14 "adjusted" – is this stratosphere-temperature adjustment?

Yes, now "adjusted" is replaced by "stratosphere-adjusted".

1:23 and throughout: I think it better to talk of a "cloud radiative effect", as is now common in the literature, rather than a "net radiative forcing" of cirrus.

We now use the term in Figures 1 and 2, e.g. We still feel that the term "net RF" is appropriate for contrail cirrus.

2:2 "heat induced" – maybe better as "changes in radiative heating"

Basically we agree, and write "Heat induced by radiation"

2:7 "covers" –> "is estimated to cover"

We agree, and changed the text accordingly.

2:26 "contrails occur mainly over land" – this could be clearer – do you mean that most flights are over land, or conditions for contrail formation are more likely over land? I think it is the first of these.

We mean that most flights are over land, and clarified this aspect in the text, accordingly.

*3:8-9 As is discussed by Hansen et al. (and I think in papers by Ponater) it needs to be clear that lambda_co2 is not a fixed number even in a single climate model, as it depends on the size of the CO2 perturbation. Hansen et al are careful to define their efficacy relative to a specific CO2 change (see their para 34 and Table 1), so that other CO2 perturbations have themselves an efficacy that departs from 1 relative to their specific case.

We agree that λ_{CO2} depends on the size of the CO_2 perturbation, and changed the text into "for a given change in CO_2".

4:1 "rating" – I didn't quite understand this word – perhaps it is "rerouting" afflicted by an automatic spell-checker?

We meant "rating" in the sense of "assessing". But now feel that this statement is no longer necessary and deleted this part for shortness and to avoid such misunderstanding.

4:18 "similar to a dust layer" – I didn't understand – mineral dust layers (if that is what is meant) can have a LW forcing.

We tried to find an example that is mainly scattering and in the SW range. We now write: "similar to a layer of small and non-absorbing particles"

5:9 And F_T is also zero in the radiative equilibrium case, I presume. If so, perhaps the C3 text should say this.

We agree and changed the text.

**5: This page needs much better structure and to establish a consistent terminology. Three cases are presented "pure radiative equilibrium", "radiative-diffusive mixing " and "radiative-convective mixing" . But sometimes different terminology is used. 8:12 refers to the "radiative case" (but all cases are radiative), Fig 10 caption refers to "radiative equilibrium without mixing" and "strong diffusive mixing", Figure 11 refers to "radiative equilibrium with zero mixing" and "uniform diffusive mixing" and then Figure 12 refers to "radiative equilibrium with zero turbulent fluxes" and "moderately strong diffusive mixing". I could go on. I hope the author will see the need to adopt a concise and consistent terminology but also to consider whether a good scientific purpose is served by presenting results for cases in almost all figures. The terminological confusion is further accentuated on page 6 by having two variants to determine the skin temperature – no separate name is given to each case, and I am frankly not sure it is necessary to even present results from both, as the zero surface turbulent heat flux case is entirely theoretical.

We see the need to adopt a concise and consistent terminology.

We now define three cases: a "radiative case" with zero turbulent fluxes, a "radiative-convective case" with turbulent mixing in unstably stratified layers and a "radiative-diffusive case" with constant diffusivity in the troposphere and zero diffusivity in the stratosphere. --- And we changed the text at several places for consistency.

*5:15 "model includes a cirrus layer" – I think it is equally important to make clear that it ONLY includes a cirrus layer – i.e. no other cloud layers are included. The paper was not clear on this point but I regard this as a serious restriction when it comes to specifically looking at the impact of contrails, and so it is important that this is kept in mind. The impact of cirrus on the surface LW and SW budget, as well as the radiative heating at cirrus cloud base (e.g. Figure 7), will be considerably affected by the lower level clouds which are missing here.

We agree on the facts. The model code is prepared to include a liquid water cloud besides a cirrus cloud, and we tested it, but we now refrain from showing further results to reduce complexity. But we do mention that other clouds are important for the quantitative results.

*5:21 It is not clear where the value for diffusivity comes from. Some earlier study? The value plays such an important role in the analysis that it has to be justified in a more rigorous way. And it is important to again acknowledge important caveats: in this case, vertical heat transport in the real atmosphere is not, for the most part, diffusive, and so what is adopted here is a convenience for the simple model. C4 5:25-29 It is a little hard to follow this – given the signs shouldn't the "max" on line 26 be a "min"?

We now discuss the value for diffusivity in more detail, The turbulent flux F_T is approximated as a function of a potential temperature gradient in the linearized form $dT/dz-\Gamma$, including the prescribed lapse rate Γ and diffusivity κ (Ramanathan and Coakley, 1978; Liou and Ou, 1983). The inclusion of Γ makes sure that an atmosphere under threshold conditions with $dT/dz = -\Gamma$ experiences zero turbulent fluxes. The diffusivity κ is set to zero in the stratosphere and to a constant κ= 100 $m^2 s^{-1}$ in the troposphere for simulation of diffusive mixing in this study. Liou and Ou (1983) used values up to 200 $m^2 s^{-1}$ to simulate cirrus in the tropical atmosphere. The diffusivity κ causes vertical mixing in the troposphere with time scales L_V^2/κ depending on vertical scales L_V of temperature changes, about 10 d for mixing over the whole troposphere ($L_V \approx$ 10 km) and about 3 h for a layer of 1 km depth. Stone (1973) estimates the effective diffusivity κ_H for horizontal mixing by large-scale eddies to be at least 10^6 $m^2 s^{-1}$. For similar time scales, the diffusivity κ for vertical mixing should be related to κ_H by the square of the ratio of vertical to horizontal length scales. The length scale ratio can be estimated from geostrophic equilibrium, $L_V/L_H \approx f^2/N^2$ where f and N are the Coriolis and the Brunt-Väisälä frequencies (Vallis, 2006). For typical mid-latitude and tropospheric values ($f= 10^{-4}$ s^{-1}, $N= 0.01$ s^{-1}) one obtains $\kappa \approx (L_V/L_H)^2 \kappa_H \approx 100$ $m^2 s^{-1}$. These are of course only order of magnitude estimates.

*6:10 Using a surface albedo of 0.3 is a very crude way of mimicking low level clouds, and of course only does so in the SW (and so the LW surface budget is more sensitive to atmospheric perturbations than it would otherwise be). It is not quite clear to me why other clouds are excluded – is it an attempt to simplify or a methodological difficulty in including them? And why 0.3? I recognise this is the planetary albedo, but a surface albedo of 0.3 does not yield a planetary albedo of 0.3, because of atmospheric absorption (pushing one way) and Rayleigh scatter (pushing the other). It would be reassuring to know what the control top-of-atmosphere radiation budget is, as this would help determine how realistic the forcings (especially the longwave) are.

The albedo and SZA were initially selected because we also wanted to study cirrus effects globally. We now concentrate on mid-latitude values similar to previous contrail studies. The basic massage of the results is unchanged. We are now even more certain that our results are robust with respect to major changes.

In addition, we learned a lot from this exercise with respect to the quasi linear behavior of the model results and the different importance of SW and LW effects, and explained this in the revised text.

*6:10 "cos(SZA)=0.25" – this surprised me too. I understand that this yields the correct incoming solar radiation at top of atmosphere, but the high zenith angle (75 degrees) will significantly bias the SW effect of contrails to be more negative – indeed it is the zenith angle close to the most negative radiative forcing, according to the excellent Schumann et al. (2012 - 10.1175/JAMC-D-11-0242.1) paper and this may significantly affect some of the section 3.2 results . In radiative convective models (such as Manabe and Wetherald) it is common to assume a cos(SZA) of 0.5 and to assume a fractional day length of 0.5, although it may be more preferable to integrate over zenith angle.

Same response as above.

*7.10 Following on from the above comment, I am now a bit further confused. In the caption of Figure 2, it refers to the daily mean at 45N on 21 June. How does this relate to the cos(SZA)=0.25? And why is a surface albedo of 0.2 used here when it is 0.3 in the text? I guess Figure 2 is trying to justify the use of the 2-stream hexagons scheme used in the radiative-convective model, but it seems to me that it is not testing it for the conditions applied in that model. I am sorry if I misunderstand. And I have a similar query about Figure 3. Since, from my understanding, the radiative-convective model does not integrate over the diurnal cycle, this plot leaves a somewhat misleading impression and I am not sure of its purpose here. My bigger question is whether the C5 choice of cos(SZA)=0.25 leads to a bias in the SW budget of contrails. Also since a cirrus optical depth at 500 nm of 0.3 (10:25) is applied in the experiments, it is not clear why a value of 0.5 is used in Figure 3.

The confusion comes from the parameters used in Meerkötter et al. (1999). We now eliminated this discrepancy in using their values.

7:19 The figures show 360-720 ppb, the text says 300-600 ppb

We agree. But the figure is now deleted, as you suggested in the next comment.

7:17 Figure 4: While it is useful for the authors to have performed this calculation, I see no reason for including it in the paper – it is a result that is over 50 years old and in my view just inflates the paper. I feel something of the same way about Figures 5 and 6, since they are referred to only in passing. The inversion in Figure 6 may be something of an artefact resulting from the exclusion of lower level clouds

Figure 4 to 6 were shown to demonstrate that the code is able to compute the convective adjustment correctly. We now feel that convective adjustment is less important for this study. Hence, these figures are no longer necessary and we eliminated the figures and the corresponding sentences.

8:6 The expression for heating rate is textbook physics and doesn't need including – I am not sure the value for the lowest level is in any case correct, if the surface pressure is really 1013 hPa (I get 0.64 K/day).

We agree (though not all readers may have your knowledge), and we now reduced details.

8:12 "radiative EQUILIBRIUM case"

We agree and changed the text.

8:13 "smaller vertical scales" – it is hard to see this when the plot is presented in linear pressure.

We agree, and now deleted this argument - it is not necessary.

8:17: I agree that the 8-13 micron window is "more transparent" than neighbouring spectral regions, but it is hardly transparent, because of water vapour continuum absorption in this region.

We agree on the physics. We now changed the text to clarify this issue.

8:19 "stratosphere" – this sentence only makes sense to me if it is the "lower stratosphere"

We agree and changed accordingly.

8:21 "rather stable" – it is unclear what measure of stability is being used in making such a statement

We agree, and this text part was eliminated for shortness.

9:23 Perhaps 2 significant figures are enough in this and later paragraphs?

We agree and changed the numbers to 2 digits.

10:5-20 The experiment described here (100% cirrus, 150% perturbation to humidity) feels very contrived and in my view was a distraction. I suggest it be removed. C6

We follow the suggestion, and remove the strong cirrus and enhanced humidity cases from the figure. The results are still mentioned.

*10:24 Why 3% given the 0.2-0.5% at 2:7? But I am concerned that the assumed cirrus amount will ultimately impact the radiative heating in the upper troposphere and hence the extent to which that region can be decoupled from the atmosphere below. In addition, I suspect that the impact is also highly dependent on the height of the cirrus as well as the assumptions about underlying clouds.

The 3% is appropriate for mid-latitudes.

The introduction now says “Contrail cirrus clouds of significant optical thickness (>0.1) are estimated to cover about 0.2 - 0.5 % of the Earth, with higher values in northern mid-latitudes”

The cirrus amount is important for convective adjustment, but not for fixed diffusivity. This is no further explained in the text

Of course, the height of the cloud layer is important as are many other parameters. But the basic message, that mixing is important is independent of the specific parameter values. See revised manuscript.

11:1 “weak turbulent mixing” – which case is this referring to? See my comment **5. If you mean zero-mixing, the text should say this.

Yes.

11:6 “only for strong” – but as I understand you have only performed the experiment for zero or strong, so there is no intermediate case? I then get further confused by the discussion of convective mixing later in the paragraph, partly for the reasons discussed above, but partly because it is not shown on Figure 11. I suspect the result is also highly sensitive to the assumed cirrus height. Comparing Figure 5 and 6, it seems clear that convective mixing is impacting the temperature profile throughout the depth of the troposphere so it confused me to say that “convective mixing is weak”

We now revised the text considerably and hope that it is now clearer.

*11:10 The discussion at 6:10 about the chosen solar zenith angle calls into question this result, and I suggest it is revisited.

The case Q_0 = 0 is no longer discussed in detail (just mentioned), to reduce unnecessary complexity.

12:18 I can see no such plot in Ponater et al. (2006) – I am sorry if I miss it. Perhaps the text should refer to Figure 2 of Ponater et al. (2005) (see also 14:25) but even there I am a bit doubtful whether the point being made is the full story;

the maximum in upper tropospheric warming may be a result of well-known moist adiabatic processes (in which a surface perturbation is amplified at upper levels via the divergence of moist adiabats with height).

We intended to refer to Ponater et al. (2005).
The maximum in the upper troposphere cannot be explained by local release of latent heat, because the mass of water that condenses during cloud formation at those low-temperature levels is small. The pattern with enhanced temperature in the contrail region is also far more pronounced than in a similar CO_2 disturbance simulation (see Figure 1 of Ponater et al. (2006)).

The 2006 paper is a conference paper, available from http://elib.dlr.de/54467/. Here we show two essential panels from the Fig 1 in that paper (with permission by Michael Ponater).

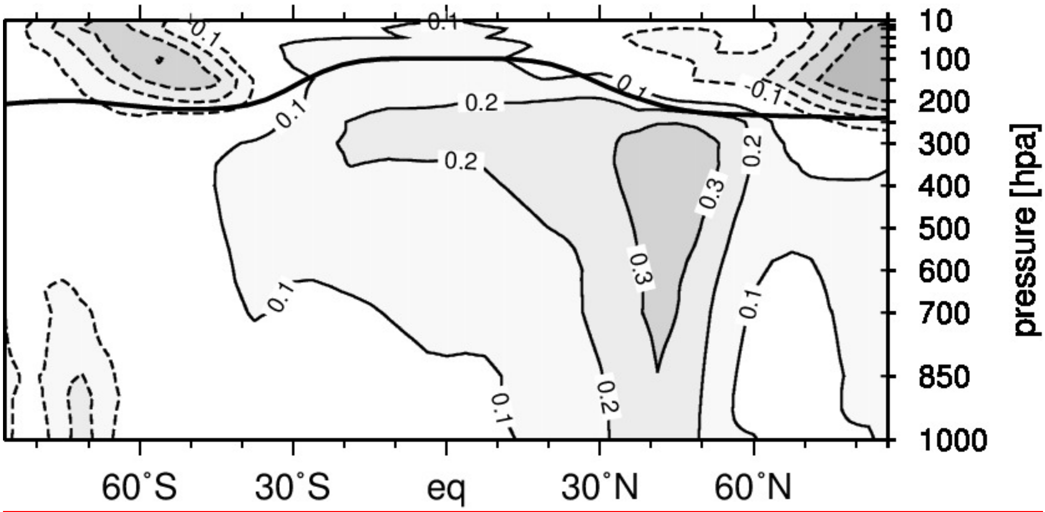

Figure a: Zonal mean temperature response (in K) caused by contrails (RF = 0.19 W m^{-2}).

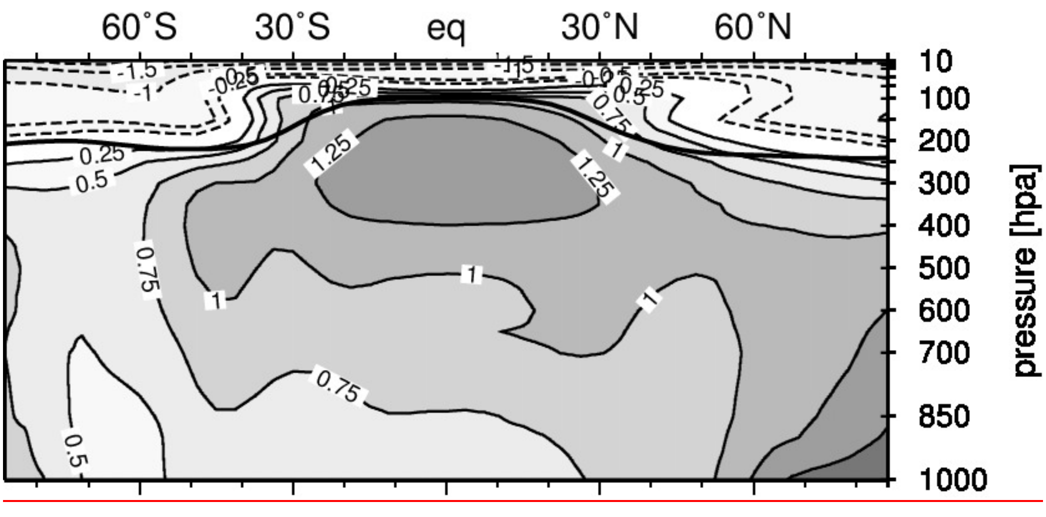

Figure b: Zonal mean temperature response (in K) caused by CO_2 disturbances (RF = 1 W m^{-2}).

13: I found this discussion rather conjectural and suggest it could be removed

We thought that the discussion should be of interest. Now we decided to remove chapter 4, and keep just a few sentences which are now in the newly formulated final Section “4 Summary, Implications and Conclusion”.

*14: Although the central idea of this paper may indeed prove to be correct, this conclusions need to draw attention to the many caveats about the simple model that is adopted and how these may impact on the final result. C7

We agree, and revised the text. The conclusions now explicitly mention the model and parameter dependence and list arguments in support.

References cited here:

Liou, K.-N., and S.-C. S. Ou: Theory of equilibrium temperatures in radiative-turbulent atmospheres, J. Atmos. Sci., 40, 214-229, 1983.

Ponater, M., V. Grewe, R. Sausen, U. Schumann, S. Pechtl, E. J. Highwood, and N. Stuber: Climate sensitivity of radiative impacts from transport systems, in: Proceedings of an International Conference on Transport, Atmosphere and Climate (TAC), edited by: Sausen, R., Blum, A., Lee, D. S., and Brüning, C., University of Manchester and DLR Oberpfaffenhofen, http://elib.dlr.de/54467/, 190-196, 2006.

Ramanathan, V., and J. A. Coakley: Climate modeling through radiative-convective models, Rev. Geophys., 16, 465-489, 1978.

Stone, P. H.: The effect of large-scale eddies on climate change, J. Atmos. Sci., 30, 521-529, 1973.

Vallis, G. K.: Atmospheric and Oceanic Fluid Dynamics, Cambridge Univ. Press, Cambridge, 2006.

Ulrich Schumann and Bernhard Mayer 22 August 2017, changed version.

Responses to Reviewer 2.

We thank the reviewer for his thoughtful and detailed review. The review comments lead to considerable changes and several improvements.

The authors investigate the extent to which top-of-atmosphere forcing from jet contrails are able to influence the surface temperature using a simple radiative-convective diffusive model. This may be phrased as the "efficacy" associated with such forcing, and the authors show that this efficacy is strongly dependent on the assumed mixing within the model.

We agree. We now add “and efficacy relative to CO_2 changes” in the abstract to follow your interpretation.

The results of this study are of interest, but they are derived from a very simplified model, and because of its simplicity I am a bit unclear on the implications of this study for Earth’s atmosphere. In particular:

We are pleased that the results are of interest.

We agree that the results are based on a simple model. That was the purpose.

As you know, our team also runs more comprehensive climate models. The problem is that such models often do not allow identifying reasons for certain results. Therefore, we looked by purpose for the most simple model we could think off to study the relative importance of mixing and radiation in clear isolation from other processes. In the conclusions, the importance of the model simplifications is now stressed. Further the abstract says: “Since the results of this study are model dependent, they should be tested with a comprehensive climate model in the future. “

1) The authors find that in the limit of weak tropospheric vertical mixing, the effect of upper tropospheric forcing like that of contrails can be to cool the surface. This seems to run counter to GCM studies of Hansen et al. (2006) and Ponater et al. (2006), which show a more constant tropospheric response, presumably because they have some vertical mixing. Does this mean that the weak vertical diffusion case in this study is simply not relevant to Earth’s atmosphere?

We agree that the limiting result cannot be guaranteed to be fully relevant for real atmospheres and we now say this in the conclusions. But as noted in the introduction recent research indicate that strong SW contributions are getting more and more realistic. Certainly, this needs further studies and this paper may trigger such studies.

2) In the mid-latitude case, convection is hardly active because the large-scale forcing Q_0 stabilises the atmosphere. But in Earth’s atmosphere, convection acts intermittently and the convective mixing is therefore underestimated by this model. Further, I think it is unreasonable to expect Q_0 to remain unchanged in response to the forcing. The thermal stratification of the midlatitudes is set by this large-scale forcing, and a change in this thermal stratification will likely have an influence on the midlatitude eddies. Is the vertical diffusion meant to be a parameterisation of these missing processes? If so, what level of vertical diffusion is relevant for Earth’s atmosphere?

The reviewer addresses important issues, which we cannot answer strictly without running far more extensive models. Our point should still be valid that mixing is important. The question whether our study gives correct quantitative result cannot be answered without further research. We now say this in the abstract and in the conclusions.

3) In the "tropical" case (Q_0 = 0; Fig. 6) the convective adjustment is controlling the lapse rate, as is the case in Earth's tropics. But here, the forcing applied is very strong: 100% Cirrus cover. In this case, the Cirrus produces an inversion in the upper troposphere, and drives a second convective cell above the tropopause. I'm not sure this is a plausible outcome of contrail forcing. What happens if the forcing is reduced to a cloud cover of 0.2-0.5%? Do you still get the same decoupling from the surface? How does this depend on the height of the forcing?

You are right that the cirrus cover is important for stabilization and we mentioned that. We now deleted this part to reduce the complexity of the paper.

4) What does Fig 11 look like with radiative and convective adjustment? we expect the CO2 response to warming to be relatively uniform in the troposphere. This is true with high diffusion, but does not seem to be true in the radiative-convective case. To me this suggests that the no diffusion limit is not relevant for the Earth.

We now show the results also for convective adjustment. This discussion is part of the discussion on model dependence. We now point out that global models often show a rather smooth profile of temperature increase in the troposphere, partly perhaps because of strong mixing on coarse grids.

My suggestions to improve the manuscript in light of these comments are as follows:

Thank you for your suggestions. We revised the paper accordingly.

- More consistent forcing levels across the experiments. In some cases the Cirrus cloud cover is set to 3%, in others 100%. Why is this the case? And how was 3% chosen? It seems much larger than the 0.2-0.5% quoted in the introduction for Contrail fraction. Does the response depend on the size of the forcing? What about the height of the forcing?

We keep less cases, with 3% cover (the 100 % case is kept in the comparison to Meerkötter et al (1999) who run the test cases with 100 % contrail cirrus cover). We discuss the importance of cirrus properties. The strength of the cirrus forcing is important mainly for convective mixing. The diffusive mixing is linear in this model and less sensitive to the contrail details. The main issue that contrails have positive RF at TOA and negative RF at the surface is robust and independent of such details.

- Some more discussion on how the results from the simple model should be interpreted. In particular, what is the level of vertical mixing relevant for Earth's atmosphere in midlatitudes and in the tropics? How does the assumption of diffusive mixing affect the results.

We agree. We now relate the diffusivity to the studies by Stone on baroclinic adjustment by large scale eddies.

- I think the study would benefit from using single-column model with a more realistic description of convection than the simple model used here (e.g., the single column model of a IPCC-class GCM). While this does not ameliorate all the problems with using a 1-D description of the atmosphere, it will ensure the convective response given the mean state will be somewhat realistic, particularly for the "tropical" case in which Q_0 is zero.

We do not follow this suggestion because we will never find a 1-d model that includes all known effects. (A future model version should be 2-dimensional and include diural and seasonal cycles – coming closer to a full climate model.) Instead, by purpose, we simplify the study even further and skip some results of variants for clarity The test against reality has to be done within comprehensive climate models. We say this in the Conclusions. However, we show and explain the robustness of the results to parameter changes.

Minor comments:

page 1: Line 8: What does "basically without climate system changes" mean? Does this refer to the dynamic heating in the model? This should be clarified here and in the other places where this statement is made in the manuscript.

The term "basically" was used since we had a model variant with fixed relative humidity. But we now skip this and simplify the paper.

page 2: line 33: Here it is argued that contrails do not behave the same as high clouds, but later the forcings you apply are described as either thin cirrus or contrails. This contradiction should be resolved

We thought that our study is relevant beyond contrail cirrus. That caused part of the complexity and apparently misleading wording. We now decided to reduce the paper to the mid-latitude case and talk about contrail cirrus only (with a short remark on generalization potentials in the Conclusions).

page 3: Line 32: I am not sure what it means to avoid warming contrails. Does this mean that one mitigation option is to move flight paths to regions in which the effects of contrails is a cooling?

Your are right in your interpretation. We now added "route changes" to clarify this question.

page 5: Line 1-10: The discussion here is very confusing. At one point it is stated that Q_0 is the sum of the divergence of F_R and F_T, but it is a bit unclear whether this statement is supposed to only apply for T = T_0 or more generally. Later it is stated that the Q_0 = 0 case is "pure radiative equilibrium", but I think this should be Q_0 = 0 and F_T = 0.

We now changed the text to avoid such misunderstandings.

page 5: line 20: I don't understand why \Gamma drops out of the equation for \Delta T, or why the contribution from \Gamma affects Q_0. Isn't Q_0 fixed? I think the equation for \Delta T should be presented for clarity.

The reference lapse rate Γ drops out for constant diffusivity. This can be seen when taking the difference of Eq. (1) for $\Delta T = T(t,z) - T_0(z)$. The corresponding equation for ΔT would make the text more lengthy without providing much insight. Basically this is of theoretical importance only. The code includes the full set of equations. Therefore, we now deleted this sentence.

page 6: line 10: Setting the cosine zenith angle to 1/4 biases the solar radiation to have a high zenith angle, this will increase the reflection from clouds and bias the results. For the global mean, one should use the insolation weighted zenith angle (Cronin 2014). But I do not see why the global mean insolation is necessarily desired. The temperature profile used is one of the mid-latitudes, so presumably that is the focus. Why not use a diurnally varying solar insolation for e.g., 45 deg?

We changed the values to mid-latitude values. The results are robust to these changes.

page 6: line 20: The radiation only boundary condition for T_skin is unphysical for cases with turbulent fluxes. Perhaps it would make more sense to use an assumed value of the surface enthalpy exchange coefficient and wind speed that are typical of Earth's surface conditions.

We decided to reduce complexity by setting the surface temperature equal to the temperature in the lowest model layer, throughout the paper. Again, a more realistic model would require further model parameters, which we want to avoid, because any parameter requires a discussion on its validity and limitation and this would make the paper just more complex without much gain and without changing in the basic conclusions.

page 8: line 32: The Hansen et al. (1997) result needs explaining. What type of model were they using? Does this indicate that the strong mixing limit is the appropriate one?

We now explain that Hansen et al. (1997) used a GCM with rather coarse resolution.

page 10: line 25: Here 3% Cirrus coverage is used, but the global cover mentioned in the introduction is 0.2-0.5%. Does the magnitude of the Cirrus cover have any effect on the results?

We use 3 % because that is representative for mid-latitudes. We now explain this.

page 11: line 9: It appears that the Cirrus drives convection above it to the tropopause. Is this likely for the forcing from Jet contrails in the next century?

It is well known that the radiative heating in a cloud layer may drive convection above it, and this is what the model simulates. This does not mean that all contrail cirrus cause convection, and we do not say that. The text got modified for avoid this misunderstanding.

Ulrich Schumann and Bernhard Mayer, 22 August 2017

Sensitivity of surface temperature to radiative forcing by contrail cirrus ~~and contrails~~ in a radiative-~~convective~~mixing model

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Abstract. ~~Earth~~Earth's surface temperature ~~changes induced~~sensitivity to radiative forcing (RF) by ~~added thin~~contrail cirrus ~~or contrails~~and the related RF efficacy relative to CO_2 are investigated ~~with a radiative-convective-diffusive~~in a one-dimensional idealized model~~, basically without climate system changes, with relaxation~~ of the ~~temperature profile by~~ atmosphere. The model includes energy transport by shortwave (SW) and longwave (LW) radiation and ~~mixing.~~by mixing in an otherwise fixed reference atmosphere (no other feedbacks). Mixing includes convective adjustment and turbulent diffusion, where the latter is related to the vertical component of mixing by large-scale eddies. The conceptual study shows that the surface temperature sensitivity to ~~cirrus~~given contrail RF depends strongly on the ~~ratio of the~~ time scales of energy transport by mixing and radiation~~, where mixing may include turbulent diffusion, convection and transports by the large-scale circulation.~~. The time scales are derived for steady layered heating (ghost-forcing) and for a transient contrail cirrus case. The radiative time scales are shortest at the surface and shorter in the troposphere than in the mid-stratosphere. ~~Heat~~Without mixing, a large part of the energy induced ~~by cirrus in~~into the upper troposphere ~~reaches the~~ by radiation due to contrails or similar disturbances gets lost to space before it can contribute to surface ~~only for strong vertical mixing. The~~ warming. Because of the different radiative forcing at the surface and at top of atmosphere (TOA) and different radiative heating rate profiles in the troposphere, the local surface-temperature sensitivity to ~~adjusted radiative forcing (RF) is larger for the shortwave (SW) than the longwave (LW) cirrus forcing. For weak mixing, cirrus may cool the surface even if the cirrus causes a positive instantaneous or~~ stratosphere-adjusted ~~radiative forcing (RF) at the tropopause. The shorter time scales near the surface indicate a potential for dominant SW surface cooling regionally where cirrus or contrails form, while weak LW warming may dominate at larger distances.~~RF is larger for SW than for LW contrail forcing. Without mixing, the surface energy budget is more important for surface warming than the TOA budget. Hence surface warming by contrails is smaller than suggested by the net RF at TOA. Under low mixing conditions, cooling by contrails cannot be excluded. This may in part explain low efficacy values for contrails found in previous global circulation model studies. Possible implications of this study are discussed. Since the results of this study are model dependent, they should be tested with a comprehensive climate model in the future.

Keywords: cirrus, contrail cirrus, radiative forcing, climate change, surface temperature, radiation transfer, mixing, efficacy

1 Introduction

~~Upper~~Contrails are similar to upper tropospheric ice clouds (cirrus) which tend to warm the troposphere by reducing outgoing longwave (LW) terrestrial radiation and cool by enhancing shortwave (SW) solar radiation backscattering (Stephens and Webster, 1981; Liou, 1986; Sinha and Shine, 1994; Chen et al., 2000; Schumann and Heymsfield, 2017). For low optical thickness, the net cloud radiative forcing (RF) from cirrus is often positive at top of the atmosphere (TOA) but negative at the surface (Ackerman et al., 1988; Stackhouse and Stephens, 1991; Fu and Liou, 1993; Jensen et al., 1994; Rossow and Zhang, 1995; Meerkötter et al., 1999; Kvalevåg and Myhre, 2007; Dietmüller et al., 2008; Lee et al., 2009b; Allan, 2011; Berry and Mace, 2014; Hong et al., 2016). For well mixed greenhouse gases, a positive RF implies a global warming (Shine et al., 1994; Hansen et al., 1997a). However, cirrus induces a radiative heat source profile which tends to warm the upper troposphere but may cool the surface (Liou, 1986). ~~Skin and near-surface air temperature changes depend on the surface heat budget which includes contributions from latent and sensible heat exchange with the atmosphere and the ground (land or ocean) in addition to the net radiation budget~~ Heat induced by radiation in the upper troposphere must be transported downwards to contribute to surface warming, e.g., by convective, baroclinic and other dynamical mixing processes ~~(Sellers et al., 1997; Lian et al., 2017)~~(Manabe and Wetherald, 1967; Stone, 1973; Vallis, 2006)~~. Heat induced in the upper troposphere must be transported downwards to contribute to surface warming, e.g. by convective mixing (Manabe and Wetherald, 1967). Hence, the question whether cirrus clouds cool or warm the Earth~~. Hence, the question whether cirrus clouds cool or warm the Earth's surface cannot be simply answered from studies of radiative flux changes alone.

The sensitivity of surface temperature to contrail cirrus changes is of relevance with respect to aviation climate impact ~~by contrails~~ (Lee et al., 2009a; Boucher et al., 2013; Lund et al., 2017). ~~Contrails are cirrus clouds induced by aircraft (Schumann and Heymsfield, 2017). Contrail cirrus of significant optical thickness (>0.1) covers about 0.2 - 0.5 % of the Earth (Minnis et al., 2013; Schumann et al., 2015; Bock and Burkhardt, 2016)~~Long-lived contrails of significant optical thickness (>0.1) are estimated to cover about 0.2 - 0.5 % of the Earth, with higher values in northern mid-latitudes (Minnis et al., 2013; Schumann et al., 2015; Bock and Burkhardt, 2016). Early studies expected a regional surface cooling from contrails (Reinking, 1968). Later, a hemispheric atmosphere warming by contrails was derived from models (Liou et al., 1990). A special report on Global Aviation of the Intergovernmental Panel on Climate Change (IPCC) (Penner et al., 1999) concluded in 1999: "Contrails tend to warm the Earth's surface, similar to high clouds". Observational evidence for contrail-warming is missing because the expected changes are small, not well correlated with contrail cover, and observed changes may have many causes (Minnis, 2005). Contrail RF contributions depend on many contrail and Earth-atmosphere system properties (Meerkötter et al., 1999; Minnis et al., 1999; Myhre and Stordal, 2001; Schumann et al., 2012). Contrails are composed of relatively small and aspherical ice particles (Gayet et al., 2012). Hence, contrails may favor the albedo cooling over the greenhouse warming effect~~, in particular for thin and high contrails and cirrus~~ (Fu and Liou, 1993; Strauss et al., 1997; Wyser and Ström, 1998; Zhang et al., 1999; Marquart et al., 2003; Wendisch et al., 2005; Markowicz and Witek, 2011; Bi and Yang, 2017). Contrail contributions to ~~RF at~~ TOA cloud forcings have been derived from observations (Schumann and Graf, 2013; Spangenberg et al., 2013; Vázquez-Navarro et al., 2015). Most traffic occurs during daytime causing contrails with higher SW fraction. The global mean positive LW and negative SW parts ~~are~~may be nearly cancelling each other with a small positive net RF at TOA. Local increases in LW fluxes below contrails are hardly measurable because tropospheric water vapor ~~effectively~~ shields the surface partly from contrail-induced LW flux changes (Kuhn, 1970). Local reductions in SW fluxes are well observable at the surface (Khvorostyanov and Sassen, 1998; Haywood et al., 2009; Weihs et al., 2015). ~~Contrails form mainly outside convective clouds in the stably stratified upper troposphere at mid-latitudes (Schumann et al., 2017), with less efficient vertical heat exchange than in the tropics (Wetherald and Manabe, 1975). Contrails occur mainly over land. It is not sure that the heat induced by contrails in the troposphere over land reaches the ocean by horizontal advection and downward mixing before getting lost to space by radiation. Contrails tend to dehydrate the upper troposphere and reduce ambient cirrus (Burkhardt and Kärcher, 2011; Schumann et al., 2015).~~ Hence, contrails may have the potential to cool (Sassen, 1997). ~~On the other hand, the contrail SW forcing may be less negative because of higher effective albedo (tropospheric system reflectance) in the extratropics than in the tropics (Stephens et al., 2015). The climate sensitivity for regional forcing at mid-latitudes may be larger than for tropical or globally uniform disturbances (Joshi et al., 2003; Shindell and Faluvegi, 2009). LW forcing may be enhanced while SW forcing may be reduced by humidity and low-level cloud changes (Kashimura et al., 2017). Hence, the equilibrium surface temperature change by contrails cannot be simply deduced from an analogy to high clouds.~~

The global mean equilibrium change of near-surface air temperature is often approximated by $\Delta T_s = \lambda$ RF as a function of the net downward flux change RF at the tropopause and a "climate sensitivity parameter" λ (Houghton et al., 1990). λ is similar to the planetary temperature sensitivity parameter λ_p to changes in solar irradiance (Stephens, 2005), $\lambda_p = [1/(4\,\sigma\,T_s^3)]\,(T_s/T_p)^3\,[dT_s/dT_p]$. Here σ is the Stefan–Boltzmann constant, T_s is the surface temperature, and T_p is the effective temperature of planetary infrared emissions, $\sigma\,T_p^4 \cong S_0\,(1\text{-}a)/4$, with solar irradiance $S_0 \cong 1360$ W m^{-2} and Earth albedo a $\cong$ 0.3. Hence, $\lambda_p \cong 0.267$ K W^{-1} m^2 for $[dT_s/dT_p] = 1$. The feedback factor $[dT_s/dT_p]$ differs from one depending on the various forcing types (Stephens, 2005; Bony et al., 2006; Stevens and Bony, 2013). Therefore, λ is not a universal constant (Forster et al., 1997; Joshi et al., 2003; Stuber et al., 2005). The "efficacy" $e= \lambda_c/\lambda_{CO2}$, i.e., the ratio of climate sensitivities λ_c for non-CO_2 disturbances and λ_{CO2} for a given change in CO_2 ~~changes~~, generally differs from one (Hansen et al., 2005). Various alternative RF definitions have been suggested to improve the link to climate sensitivity (Boucher et al., 2013; Myhre et al., 2013). The instantaneous RF_i is the RF for a fixed atmosphere. The adjusted RF_a is the RF after thermal relaxation of the stratosphere to the disturbance (Houghton et al., 1990; Stuber et al., 2001). The effective RF_s is the RF after adjustment of the atmosphere to disturbances for constant (ocean) surface temperature (Rotstayn and Penner, 2001; Hansen et al., 2002; Shine et al., 2003). ~~Temperature profile disturbances within the atmosphere relax by thermal relaxation with time scales~~ ~~t_R~~

~~which are, as we will further discuss below, of order hours to months depending, among others, on altitude, vertical disturbance scales, and mixing (Manabe and Strickler, 1964; Zhu, 1993). Because of large ocean heat capacity and efficient heat exchange between ocean and atmosphere, the relaxation times scales are far smaller than the time scales for reaching climate equilibrium (Hansen et al., 1981).~~

Since air traffic is projected to continue to increase for many decades, it is important to know the climate impact of contrails accurately (Lee et al., 2009a). ~~One-dimensional (Strauss et al., 1997) and two-dimensional radiative-convective models (Liou et al., 1990) showed that contrails may have significant climate impacts. The hope was that three-dimensional global circulation atmosphere/ocean models with a suitable contrail model provide reliable estimates of the climate impact from contrails (Ponater et al., 1996). Various models to represent contrail cirrus in~~Various models to represent contrails in three-dimensional atmospheric global circulation models have been developed ~~(~~(Ponater et al., 1996; Rind et al., 2000; Ponater et al., 2002; Marquart et al., 2003; Rap et al., 2010b; Burkhardt and Kärcher, 2011; Jacobson et al., 2011; Olivié et al., 2012; Chen and Gettelman, 2013; Schumann et al., 2015~~)~~; Bock and Burkhardt, 2016), with different ~~treatment~~treatments of traffic, subgrid scale contrail formation and optical properties. Some of these models were run with atmosphere-ocean coupling (Rind et al., 2000; Ponater et al., 2005; Rap et al., 2010a; Huszar et al., 2013; Jacobson et al., 2013). ~~All these model studies suggest a mean global warming from contrails.~~ The contrail climate effects are expensive to compute because they are small compared to the interannual variability ("climate noise") in climate models (Ponater et al., 1996; Hansen et al., 1997b), so most studies used by factor 10 to 100 increased disturbances. All these model studies suggest a mean global warming from contrails. The contrail efficacy has been computed in a few studies, with results varying from 0.3 to 1~~ for not fully explained reasons~~ (Hansen et al., 2005; Ponater et al., 2005; Rap et al., 2010a). ~~Avoiding warming and enhancing cooling contrails is considered as a potential concept to mitigate aviation climate impact if such rating is possible (Schumann et al., 2011; Grewe et al., 2017). Hence, an improved understanding of climate sensitivity to contrail cirrus is urgently needed.~~

The classical RF concept assumes that the surface temperature response follows the TOA energy budget change (Schneider and Dickinson, 1974; Dickinson, 1982). This requires that the energy induced by the disturbance gets well mixed globally within the troposphere and down to the surface. In order to be effective for a long-term ocean warming, the mixing has to occur at time scales fast compared to the time scale of heat loss from the atmosphere to space by radiation. Temperature profile disturbances within the atmosphere relax by thermal relaxation with time scales which are of order hours to months depending, among others, on altitude, vertical disturbance scales, and mixing (Manabe and Strickler, 1964; Zhu, 1993). Mainly because of denser air traffic, most contrails occur at mid-latitudes. At mid-latitudes, mixing is mainly driven by baroclinic instability in the stably stratified, rotating atmosphere, also depending on moisture, leading to large-scale eddies transporting heat from the tropics poleward and upward (Stone, 1973). The baroclinic mixing occurs at time scales of several days (Vallis, 2006). Hence, mixing at mid-latitudes is slower than in convective parts of the tropics where deep convection in clouds causes fast vertical heat transport (Wetherald and Manabe, 1975). Again because of denser traffic, contrails occur mainly over land. It is not sure that the heat induced radiatively by contrails in the troposphere over land reaches the ocean by horizontal advection and downward mixing before getting lost to space by radiation.

In this conceptual study, we investigate changes in temperature from additional thin cirrus ~~or~~ (contrails) at mid-latitudes in a radiative-~~convective model.~~mixing model where the vertical mixing may result from deep convection, from the large-scale circulation, and from turbulent diffusion. For better understanding~~ of fast adjustment processes~~, the model is run without climate system changes (“feedbacks~~”)~~”; Manabe and Wetherald (1967)) except thermal relaxation by radiation and mixing. The model is run with highly idealized surface conditions (to ~~reduce~~minimize the number of free parameters), including constant temperature and zero net vertical heat flux at the surface (“adiabatic surface”) as bounding extremes. Instead of investigating the approach to equilibrium with ocean coupling, we simulate the equilibrium atmosphere without heat exchange to an underlying compartment. The disturbances considered are small and, hence, change the reference atmosphere only slightly. For this reason the model is run with fixed dynamical heating, simulating the heat sources, e.g., from horizontal heat advection, as required for a steady-state reference atmosphere (Strauss et al., 1997). The optical properties of cirrus are essential for its radiative forcing (Fu, 1996; Myhre et al., 2009; Yang et al., 2015), but for this study, the cirrus is just a source of SW and LW radiation flux-profile changes with cloud-radiation interaction details of secondary relevance. Also, aerosol effects are not included in this study. The method is described in Section 2. Section 3 presents the results. Section 3.1 shows the responses of an idealized atmosphere to prescribed heating, so-called “ghost forcing”. This part will point out the importance of the vertical distribution of the radiative heat sources and vertical mixing. The thermal response to ~~an added thin cirrus layer, typical for~~a prescribed contrail cirrus~~,~~ layer is studied in Section 3.2. We separate the temperature responses to SW and LW radiative disturbances by ~~cirrus~~contrails and refer correspondingly to “SW cirrus” (similar to ~~a dust~~ layer of small and non-absorbing particles) and “LW cirrus” (similar to a strong greenhouse gas layer). For constant atmosphere, the sum of SW and LW RF from these cirrus versions is the same as the net RF~~ from “normal” cirrus.~~. This part will show different temperature responses to SW and LW radiative forcing. A study of thermal relaxation times for ~~cirrus will show up some consequences of temporally and spatially variable cirrus. For comparison and for computation of efficacies for cirrus relative to CO_2, we run the same simple model for changed CO_2. Section 4 discusses implications of the height-dependent thermal relaxation time scales for global warming from regional cirrus clouds, with SW and LW effects getting advected over different spatial scales. Section 4 also discusses the temperature response to cirrus with some climate system changes (feedbacks), taking the model with adjusted humidity (Manabe and Wetherald, 1967) as an example for temperature-mediated system changes. Here we show that SW and LW efficacies differ not only for the stratosphere-adjusted RF but also for the effective RF. Finally, Section 5 summarizes the findings and presents conclusions~~contrail cirrus will show up some consequences of temporally and spatially variable contrails. For comparison and for computation of efficacies for contrails relative to CO_2, we run the same model for changed CO_2. Finally, Section 4 summarizes the approach, the results, and the limitations, and mentions some implications.

2 Radiative-convective-diffusive mixing model

This study uses a one-dimensional radiative-~~convective-diffusive~~mixing model of the atmosphere with prescribed composition and clouds, following traditional approaches (Möller and Manabe, 1961; Manabe and Strickler, 1964) with turbulent fluxes as in Ramanathan and Coakley (1978).

The model is integrated step-wise in time until steady state. It computes the temperature profile T(z,t) versus altitude z and time t as induced by radiative and turbulent heat transports, based on the heat budget:

$$\cancel{\rho c_p \frac{\partial T}{\partial t} = -\frac{\partial F_R}{\partial z} - \frac{\partial F_T}{\partial z} + Q_o,} \qquad \rho c_p \frac{\partial T}{\partial t} = -\frac{\partial F_R}{\partial z} - \frac{\partial F_T}{\partial z} + Q_o,$$

$$\cancel{F_R = F_{SW}^{up} - F_{SW}^{dn} + F_{LW}^{up} - F_{LW}^{dn}, F_T = -\rho c_p \kappa \left(\frac{\partial T}{\partial z} + \Gamma \right)}$$

$$F_R = F_{SW}^{up} - F_{SW}^{dn} + F_{LW}^{up} - F_{LW}^{dn}, F_T = -\rho c_p \kappa \left(\frac{\partial T}{\partial z} + \Gamma \right). \qquad (1)$$

Here, ρ and c_p are air density and isobaric specific heat capacity, ~~F~~F_R is the net radiative flux (sum of upward and downward SW and LW fluxes), F_T is the turbulent heat flux, Γ is a prescribed threshold lapse rate, and κ=κ(t,z) is a turbulent diffusivity selected to approximate diffusive mixing (constant κ) or convective adjustment (large κ in case of unstable stratification), as explained below. For contrails and for other small disturbances we compute the temperature change profile $\Delta T(t,z) = T(t,z)-T_0(z)$ in a given reference atmosphere with temperature profile $T_0(z)$, i.e., we run the model with "fixed dynamical heating" Q_0. Here, $Q_0 = \partial(F_R + F_T)/\partial z$ is the divergence of the total ~~fluxes $F_R + F_T$, so that $\partial T/\partial t=0$,~~flux for $T = T_0$, so that the undisturbed reference atmosphere is steady. Fixed dynamical heating is commonly used for stratospheric adjustment (Ramanathan and Dickinson, 1979; Forster et al., 1997; Myhre et al., 1998)~~ but used here also for tropospheric adjustments of the~~. Here we use fixed dynamical heating to study the atmosphere response for a given reference atmosphere to small disturbances (Strauss et al., 1997)~~. Cases with pure radiative equilibrium ($Q_0 = 0$) are discussed also.~~.

The radiative flux F_R is computed with an efficient delta-Eddington two-stream solver using libRadtran (Mayer and Kylling, 2005; Emde et al., 2016)~~.~~, which is a common solver for climate model applications. Tests with the more accurate discrete ordinate solver DISORT (Stamnes et al., 1998) show flux differences relative to the two-stream solver of ~~the order~~about 10 %, ~~but DISORT takes far more computing time~~mainly in the LW range. Radiation absorption by gases (H_2O, CO_2, O_3, etc.) is calculated with correlated-k distributions for SW (~0.2 - 4 µm) and LW radiation (4 - 70 µm) from Fu and Liou (1992). An alternative SW absorption model from Kato et al. (1999) induces flux differences small compared to those between the two solvers. The model includes a cirrus layer of hexagonal ice crystals with optical properties from Fu (1996) and Fu et al. (1998).

The turbulent flux F_T is approximated as a function of ~~the~~a linearized potential temperature gradient $dT/dz-\Gamma$ including the prescribed lapse rate ~~Γ~~Γ, and diffusivity κ ~~(Ramanathan and Coakley, 1978; Liou and Ou, 1983)~~(Ramanathan and Coakley, 1978; Liou and Ou, 1983). ~~Γ is included to make~~The inclusion of Γ makes sure that an atmosphere under threshold conditions with $dT/dz = -\Gamma$ experiences zero turbulent fluxes. ~~The added Γ drops out in the equations for ΔT for fixed dynamical heating because the contribution from Γ affects also Q_θ.~~ The diffusivity κ is set to zero in the stratosphere and to a constant $\kappa = 100\ m^2\ s^{-1}$ in the troposphere for simulation of diffusive mixing in this study. ~~This value turns out to cause strong vertical mixing in the troposphere with time scales h^2/κ of the order of a few days depending on vertical scales h of temperature changes and surface boundary condition.~~ Liou and Ou (1983) used values up to 200 $m^2\ s^{-1}$ to simulate cirrus in the tropical atmosphere. The diffusivity κ causes vertical mixing in the troposphere with time scales L_v^2/κ depending on vertical scales L_v of temperature changes, about 10 d for mixing over the whole troposphere ($L_v \approx 10$ km) and about 3 h for a layer of 1 km depth. Stone (1973) estimates the effective diffusivity κ_H for horizontal mixing by large-scale eddies to be at least $10^6\ m^2\ s^{-1}$. For similar time scales, the diffusivity κ for vertical mixing should be related to κ_H by the square of the ratio of vertical to horizontal length scales. The length scale ratio can be estimated from geostrophic equilibrium, $L_v/L_H \approx f/N$ where f and N are the Coriolis and the Brunt-Väisälä frequencies (Vallis, 2006). For typical mid-latitude and tropospheric values ($f = 10^{-4}\ s^{-1}$, $N = 0.01\ s^{-1}$) one obtains $\kappa \approx (L_V/L_H)^2\ \kappa_H \approx 100\ m^2\ s^{-1}$. These are, of course, order of magnitude estimates.

Various methods have been used in the past for "convective adjustment", i.e., enforcement of the lapse rate below a given threshold of, e.g., $\Gamma = 6.5\ K\ km^{-1}$ (Manabe and Strickler, 1964; Ramanathan and Coakley, 1978). Here, we increase the diffusivities by the factor $100\ (2/\pi)\ atan(\gamma)$, with $\gamma = max[0$~~$,(\Gamma+$~~$,-(\Gamma+dT/dz)/\Gamma_t]$, allowing for a small departure of $-dT/dz$ from the threshold lapse rate Γ by setting Γ_t to 0.1 $K\ km^{-1}$. This causes rapid convective adjustment at timescales shorter than one time step (6 h) and avoids spurious numerical oscillations from the on/off behavior of convection near threshold conditions. The method provides a well-defined turbulent flux, avoids iterations, is numerically stable, and conserves thermal energy.

The numerical scheme uses a non-uniform grid in z with model TOA at 60 km with 100 grid cells vertically. High vertical resolution is necessary to resolve the local flux changes caused by thin ~~cirrus~~cloud layers. The lowest layer is centered at 25 m, the highest at 57.5 km, about 0.3 hPa; the grid spacing is $\Delta z = 250$ m between 0.25 and 19 km height. The radiative solver gets the air temperature and composition at grid centers together with the ~~skin~~ surface temperature as input and returns the fluxes at the grid cell boundaries as output. This staggering avoids 2-Δz-wave artefacts. Diffusive fluxes are computed implicitly with a tridiagonal Gaussian solver based on the temperatures at the next time step. Pressure is recomputed after each change in temperature as a function of altitude for air as ideal gas assuming hydrostatic equilibrium for given gravitational acceleration and surface pressure (1013 hPa). The tropopause is defined~~, as common in meteorology,~~ by the lowest grid interface with $dT/dz > -2\ K\ km^{-1}$.

Initial conditions prescribe temperature and composition profiles for the mid-latitude summer standard atmosphere without aerosols (Anderson et al., 1986), see Figure 1. ~~The humidity profile is kept constant unless noted otherwise.~~ Surface albedo (A = 0.~~3) is selected to mimic an average low level cloud cover, and the~~ 2), solar zenith angle (cos(SZA) = 0.~~25) is set such that the downward~~6, SZA=53°) and daytime fraction (0.64) are selected for mid-latitude summer conditions similar to other contrail studies (Meerkötter et al., 1999; Myhre et al., 2009). The 24-h mean TOA fluxes for these conditions are 525, 101 and 298 W m^{-2} for incident solar~~ direct~~, reflected solar, and outgoing longwave radiation~~ equals 1/4 of the solar irradiance as~~, respectively. The dynamical heating Q_0 required to keep the mid-latitude summer atmosphere at steady state is shown in ~~the global mean.~~ Figure 1.

Boundary conditions prescribe either fixed ~~(skin)~~ surface temperature or an adiabatic boundary. An adiabatic boundary is implemented by setting $F_R+F_T= 0$ at the surface. This flux is used when computing the heating rate in the lowest model layer. An adiabatic surface implies zero surface heat capacity and zero total flux between the atmosphere above and the compartment below the surface. This condition also simulates an atmosphere in thermal equilibrium with the lower compartment (ocean, ice, etc.). ~~We consider two variants to determine~~In this study, the ~~skin~~surface temperature ~~T_{skin} at the adiabatic surface. T_{skin} is either~~is set equal to the air temperature T_s in the lowest model layer, implying rapid ~~mixing~~energy exchange between the surface and the lowest air layer~~, or T_{skin} is determined from the surface energy budget for given surface albedo A and unit surface emissivity, $\sigma T_{skin}^4 = F_{SW}^{dn}(1-A)+F_{LW}^{dn}$, implying zero turbulent fluxes at the surface~~. The code runs stably with 6-h time steps for all applications in this paper.

The atmosphere ~~responses~~responds to the radiative heating with changes of temperature~~ and of the related fluxes~~, see Eq. (1), until the sum of the changed radiative and turbulent fluxes approach a vertically constant value. For constant surface temperature the fluxes stay non-zero. ~~The fluxes are assumed to be positive for z vertically upwards. Positive upward fluxes imply a cooling, negative a warming of the surface.~~ Over an adiabatic surface, the fluxes approach zero at all heights. During integration, we monitor the net vertical flux at all relevant altitudes (during stratospheric adjustment only in the stratosphere). The integration is performed until the maximum deviation of the flux values from the mean at all these altitudes is <0.3 % of the maximum instantaneous flux value. ~~Approach to equilibrium is accelerated, during the first half of time steps, by adding, e.g., 5 times the mean heating rates in the troposphere and stratosphere to the temperature changes in the respective layers. Here, the mean heating rates result from the differences between the fluxes at top and bottom of the layer divided by the layer heat capacity. With this method, radiative equilibrium is reached within the given deviation with less than 640 time steps (160 d).~~

RF is computed from the difference between the net total fluxes~~ at the tropopause (TP)~~ in model solutions with and without the disturbance. The sign of RF is defined such that positive values imply a warming of the Earth-troposphere system. ~~For fixed dynamical heating, the model solution without disturbance is given by the steady-state initial conditions.~~ The instantaneous (i), stratospheric adjusted (a), and the effective (s) forcing is computed from three model runs with different boundary conditions. RF_i is the flux change for fixed atmosphere; it varies with height. RF_a is the flux change at the ~~TP~~tropopause after the stratosphere temperature has adjusted to the disturbance for fixed troposphere; it is constant throughout the stratosphere. ~~RF_i and RF_a are computed for fixed skin surface temperature.~~ The effective RF_s is the flux change at the ~~TP~~tropopause after reaching equilibrium in the entire atmosphere with fixed T_s. Here, the total flux is vertically constant. Finally the equilibrium response is computed for an adiabatic surface for which the total flux change is zero at all levels.

The method has been tested with the mentioned alternative solvers and molecular absorption models by comparison of the daily mean and the time dependent instantaneous SW and LW RF values of a ~~cirrus~~contrail layer with results from earlier studies (Meerkötter et al., 1999); see Figure 2 and Figure 3. ~~The dynamical heating Q_0 required to keep the mid-latitude summer atmosphere at steady state is shown in Figure 1. On average, the heating rate from Q_0 is 1.39 K d^{-1} in the troposphere and -0.062 K d^{-1} in the stratosphere. These values are similar to the net heating rates presented in fig. 22 of Manabe and Möller (1961). For zero dynamical heating, the code reproduces the approach to pure radiative equilibrium in the atmosphere (Manabe and Strickler, 1964), see Figure 4. Because of strong variations of the heating rate with altitude, the transient solution tends to form temperature kinks in the lower stratosphere. These kinks disappear slowly when reaching equilibrium because of low energy exchange by radiation between neighboring layers, mainly in the 15-µm CO_2-band in regions with low H_2O and O_3 concentrations (Plass, 1956). Figure 4 also shows that the model simulates convective adjustment similar to Manabe and Strickler (1964), which illustrates the known importance of vertical mixing for the temperature profile.~~For zero dynamical heating, the code reproduces the approach to radiative equilibrium in the atmosphere as in Manabe and Strickler (1964). For a doubled CO_2 mixing ratio (from 300 to 600 µmol mol^{-1}), the model computes a temperature change of 1.1 K without feedbacks, similar to previous results (Hansen et al., 1981). ~~The radiative-convective equilibrium solutions with a cirrus layer for zero forcing Q_0 are shown in Figure 5 and Figure 6. These results are qualitatively similar to those presented below for deviations from the mid-latitude summer atmosphere. Of course, the mid-latitude summer atmosphere is far less convective than the free radiative equilibrium atmosphere.~~

3 Results and Discussions

3.1 Temperature response to prescribed heating at various altitude levels

In order to understand air and surface temperature responses to heating at various altitudes, we follow the "ghost" forcing concept of Hansen et al. (1997a). The ghost forcing is a prescribed additive flux change causing a constant heating rate in an altitude interval. The heating causes temperature changes until reaching equilibrium in which the changed fluxes balance the ghost forcing. ~~The model is run for fixed climate system except changing temperature and mixing. In contrast to a forcing by an added cloud or by changed air composition, the ghost forcing does not change the radiative properties of the atmosphere except by temperature changes.~~

Eleven simulations are performed with a prescribed flux change of 1 W m^{-2}. One simulation is run for a flux change in the lowest model layer above the surface, and ten for flux changes in subsequent 100-hPa pressure intervals between the surface and TOA. The imposed change in net flux is zero at the surface, without direct impact on surface heating, and decreases linearly from 0 to -1 W m^{-2} within the heated atmosphere interval. Above the heated layer, the flux change is constant ~~reflecting a change of the heat budget between the surface and TOA,~~ so that RF_i = 1 W m^{-2} at TOA. ~~For an atmosphere in hydrostatic equilibrium with dp = -ρ g dz, the ghost forcing causes a~~ Because of equal masses, the heating rate ~~(rate of temperature change)~~ H= ~~$(\partial T/\partial t)_R$ =~~ g $(\partial F_r/\partial p)/c_p$ ~~. Here, H~~= 0.~~0833~~083 K d^{-1} is constant in the respective 100-hPa intervals, ~~and~~but 0.~~825~~24 K d^{-1} in the thinner lowest layer for ~~the~~ surface ghost forcing. Figure 4 shows, for example, the heating profile for forcing between 600 - 700 hPa. Figure 5 shows the initial and final flux profiles versus height for ~~these cases.~~the disturbances considered in this paper. We find that the flux in equilibrium over a constant ~~surface -~~temperature surface is in between the initial instantaneous flux values at the ~~TP~~tropopause and at the surface.

Figure 6 shows the steady-state temperature profiles versus pressure-altitude in response to the 11 ghost forcings and for three different versions of vertical mixing~~. In the~~; a "radiative case" with zero turbulent fluxes, a "radiative-convective case" with radiative transports and turbulent mixing in unstably stratified layers, and a "radiative-diffusive case" with radiative transport and mixing by constant diffusivity in the troposphere and zero diffusivity in the stratosphere. In the radiative case, the temperature change profiles are similar to vertically smoothed heating rates~~. The profiles follow the local heating with vertical scales that are the smaller, the higher the effective optical depth for infrared~~ Here, radiation causes the energy exchange between neighboring layers and between the air layers and the surface (Stephens, 1984; Goody and Yung, 1989). ~~Radiation causes energy exchange between neighboring layers and between the air layers and the surface.~~ The atmosphere and the surface also emit energy directly to space. Even for ~~heating~~ghost forcing at the surface, the lowest air layer gets warmer than the surface because the warm black surface emits radiation to space more efficiently in the partially transparent thermal infrared window between 8 and 13 μm wavelengths while the air ~~layer emission~~ layer's emissivity is ~~weak~~lower in this spectral range. ~~Because of~~In the lower ~~emissivity and lower temperature~~stratosphere, the temperature increase required to balance the ghost forcing is far higher ~~in the stratosphere~~ than in the troposphere~~.~~ because of lower emissivity and lower temperature. Turbulent ~~vertical~~ mixing smoothes the profiles further, as expected. ~~Convective mixing is rather weak for this case because the~~ Because of stable stratification in the mid-latitude ~~summer~~reference atmosphere ~~is rather stable compared to the tropics, so that convection~~ , convective mixing occurs only in the upper troposphere where the ghost heating ~~causes~~is strong enough to cause local instability.

Figure 7~~a shows the surface temperature change ΔT_s as a function of the height of the heated layer. ΔT_s is, of course, maximum for ghost forcing directly at the surface. Its value depends on details of the surface boundary condition. Here we show results assuming perfect mixing between the surface and the lowest air layer with equal skin and air surface temperatures, ΔT_s = 0.371 K. In the alternative, for T_{skin} computed from the local radiation budget, T_{skin} is far higher (by about 13 K for the given albedo and SZA, which is a realistic magnitude (Lian et al., 2017)) and emits energy more efficiently, so that the skin temperature change induced by ghost forcing is smaller, ΔT_{skin} = 0.300 K, and the air surface temperature change is larger, ΔT_s = 0.491 K. Without diffusive mixing in the troposphere (black circles), ΔT_s decreases with the height of the heated layer. For strong tropospheric mixing (κ =100 m^2 s^{-1}, red symbols), ΔT_s is 0.260 K for surface ghost~~

~~forcing, and this value stays close to constant within the whole troposphere.~~a shows the surface temperature change ΔT_s as a function of the height of the heated layer. As expected, ΔT_s is maximum for ghost forcing directly at the surface, $\Delta T_s = 0.37$ K. Above the surface, ΔT_s decreases with the height of the heated layer. So, the ghost forcing efficiency in heating the surface by radiation transfer decreases with layer height. For diffusive mixing, ΔT_s is smaller (0.26 K) and stays close to constant within the whole troposphere. For comparison, Hansen et al. (1997a) (their Table 4 and Fig. ~~8~~8a) use a~~)~~ coarse-resolution global circulation model and report a vertically nearly constant ΔT_s for fixed clouds, with ΔT_s = 0.~~288~~29 K when normalized to the same forcing. Apparently their model simulated strong vertical mixing. Small differences were to be expected because of, e.g., different atmospheres.

Figure 7c shows the thermal relaxation time scale $t_R = \Delta T/H$ (in units of days) computed from the steady-state layer-mean temperature change ΔT in the heated layers at various levels and the ~~given~~corresponding heating rate H. For the radiative ~~equilibrium with zero turbulent fluxes~~case, t_R is 0.45 d near the surface (and smaller for thinner surface air layers), 6.6 d in the first 100 hPa layer, 11 d in the upper troposphere, 30 d in the ~~TP~~tropopause region between 100 and 200 hPa, and 23~~.5~~ d in the top 100-hPa layer. For layers with 200 hPa depth instead of 100 hPa, the heating response is smoother, causing about 50 % larger time scales. ~~Hence, the sensitivity to layer depth is less than linear~~Hence, as expected (Goody and Yung, 1989)~~.~~, the sensitivity to layer depth is less than linear. Radiation causes nonlocal energy transfer, different from diffusion processes for which the sensitivity to layer depth would be quadratic. The smaller time scales in the lowest layers are again a consequence of effective radiation emission via the surface. The relaxation times in the highest layer are lower than in the second highest layer, because of stronger heat loss from the middle atmosphere to space (Zhu, 1993). ~~Turbulence causes additional~~As expected, mixing ~~reducing~~reduces the layer warming and the related time scales. Mixing in the troposphere also reduces stratospheric time scales by enhanced heat exchange ~~between air layers near the tropopause by radiation, heat exchange~~ with and within the troposphere~~ by mixing~~, and enhanced heat loss from the surface to space. With ~~strong tropospheric vertical~~the diffusive mixing, the thermal relaxation times for heating in the troposphere approach a low and vertically constant value of about 3.2 d. For an atmosphere in which the adiabatic surface is replaced by a constant temperature surface, the time scale t_R is zero at the surface; t_R reduces by 34 % in the first 100-hPa layer, and by 12 % in the second layer, with smaller changes at higher levels. In ~~this~~the diffusive case, because of combined transport by radiation and mixing, heat has a lower residence time than a passive tracer with similar source location and constant concentration at the ~~Earth~~Earth's surface. ~~Passive~~For comparison, passive aircraft emissions may well exceed one month atmospheric residence time when emitted into the lower stratosphere (Forster et al., 2003) but reach ground within less than about a week when emitted in the mid troposphere (Danilin et al., 1998).

Figure 7b and e show the adjusted and effective RF_a and RF_s versus the height of the heated layer. RF_a equals RF_i = 1 W m^{-2}, regardless of the layer height as long as the heated layer is fully below the ~~TP~~tropopause (Hansen et al., 1997a). The ratio RF_s/RF_i measures the fraction of heat that ~~continues to warm~~warms the compartment below the surface after the air temperature has adjusted to the induced heat disturbance. RF_s/RF_i is largest for heating near the surface: 0.~~804 in the case~~ 80 without diffusive mixing. Hence, after fast adjustment, when the troposphere has reached its higher steady-state temperature, about 80 % of the input heat heats the compartment below the ~~Earth~~Earth's surface (e.g., ocean) and 20 % of the heat radiates out to space~~ when the troposphere has reached its higher steady-state temperature.~~. For heating near the ~~TP~~tropopause, about 95 % of the heat leaves to space. For strong vertical mixing, RF_s/RF_i is about 60 % and vertically nearly uniform. Hence, even with strong mixing, ~40 % of the ghost heating radiates directly to space.

Finally, Figure 7d and f show λ_a and λ_s, the sensitivity parameters of ΔT_s to RF_a and RF_s. For heating at the surface, λ_a = 0.~~371~~37 K W^{-1} m^2~~ based on equal skin and air surface temperature. It would be 0.291 K W^{-1} m^2 and, hence, closer to~~. This value is larger than the planetary sensitivity (0.~~267 K W^{-1} m^2 for [dT_s/dT_p] =1) if based on skin surface temperature~~27 K W^{-1} m^2, without ~~surface mixing.~~feedbacks) because the atmosphere reduces heat losses from the surface. Without diffusive mixing~~ (black circles),~~, the ~~value~~values of λ_a ~~decreases~~decrease strongly with height, because heating at higher levels is less efficient in radiative surface warming. With ~~strong ~~diffusive mixing~~ (red symbols),~~, λ_a approaches a constant because the heating is distributed quickly over the troposphere regardless of the layer height. ~~The~~In contrast, the value of λ_s is close to a constant because RF_s already accounts for the fast temperature profile adjustment. Therefore, ~~RF_s is a better measure for surface temperature change than RF_a.~~ as expected (Shine et al., 2003), RF_s is a better measure for surface temperature change than RF_a.

~~The response to ghost forcing characterizes the thermal response for a fixed atmosphere. In addition to mixing, the thermal response depends, of course, on the temperature and composition of the atmosphere. Large changes result from added clouds or from changes in air composition such as humidity. Figure 11a (cyan symbols) shows that ghost forcing below the cloud causes a larger surface temperature change when the reference atmosphere is covered with 100 % cirrus of visible optical thickness τ = 3 at 10 to 11 km altitude. The cloud reduces the heat loss to space. The cirrus cloud must be quite thick to effectively shield the lower troposphere from radiative heat losses. Note that the infrared absorption optical thickness is typically only half of the visible optical thickness (Garnier et al., 2012). Hence, even for 100 % cover, the solar optical thickness must exceed about 2 to cause a notable reduction on radiative heat losses from the troposphere to space. The plot also shows that increasing the humidity profile to 150 % of the initial value uniformly at all altitudes in the reference atmosphere reduces surface warming by ghost forcing slightly. A uniformly higher humidity in the atmosphere enhances the infrared layer emissivity, causing stronger local cooling from a ghost layer to space; it also increases the optical thickness between the layer and the surface, reducing surface temperature changes. This is no contradiction to the fact that increases in stratospheric water vapor (and CO_2) act to cool the stratosphere but to warm the troposphere (Shine and Sinha, 1991; Solomon et al., 2010). We applied the code also for the tropical standard atmosphere (Anderson et al., 1986). In the more humid tropics with higher and colder tropopause, the relaxation time scales are about 20 % smaller than at mid-latitudes. For an atmosphere with doubled CO_2, the changes are qualitatively similar to increased H_2O, but of smaller magnitude. High and thick clouds are far more efficient in changing the radiative relaxation time scales in the troposphere than added H_2O or CO_2.~~

Since ghost forcings change the temperature, they affect LW radiation. The changes depend solely on the temperature profile of the reference atmosphere and the infrared optical properties of the atmosphere and the surface. The solar irradiance is unimportant for fixed dynamical heating. The model response is quasi linear in the magnitude of the disturbances for fixed mixing properties as long as the temperature changes are small compared to absolute temperature. To illustrate the quasi linearity, we tested the model with ghost forcing increased from 1 W m^{-2} to 4 W m^{-2}. The values of t_R for ghost forcing at the surface. e.g., get reduced by up to 0.5 % to 5.7 % for this change, for the three cases, with largest changes for the radiative-convective case. For λ_s, the values range from 0.6 % to 1.2%. Further tests have shown that the basic altitude dependence in the sensitivity to ghost forcing exists also for zero dynamical heating.

We applied the model also for atmospheres with an additional cirrus layer in the upper troposphere, with increased humidity, with increased absolute humidity keeping the relative humidity constant, with increased CO_2, and for other standard atmospheres. All these changes cause changed t_R, RF_s, and temperature sensitivity λ_s values. Clouds of sufficient optical depth above the heated layers reduce the heat loss to space notably. A uniformly higher humidity in the atmosphere enhances the infrared layer emissivity, causing stronger local cooling from a ghost layer to space; it also increases the optical thickness between the layer and the surface, reducing surface temperature changes. This is no contradiction to the fact that increases in stratospheric water vapor (and CO_2) act to cool the stratosphere but to warm the troposphere (Manabe and Wetherald, 1967; Shine and Sinha, 1991; Solomon et al., 2010). In the more humid tropics with higher and colder tropopause, the relaxation time scales are about 20 % smaller than at mid-latitudes. The response to changes for fixed relative humidity helps to understand climate change feedbacks (Manabe and Wetherald, 1967), but requires a more extensive model to be realistic. For an atmosphere with doubled CO_2, the changes are qualitatively similar to increased H_2O, but of smaller magnitude. High and thick clouds are far more efficient in changing the radiative relaxation time scales in the troposphere than added H_2O or CO_2.

3.2 ~~Cirrus~~ Contrail cirrus in comparison to CO_2

In this section we consider the temperature changes induced by a contrail cirrus example, a thin homogenous cirrus layer at 10 to 11 km altitude, with 3 % coverage (typical for mid-latitude contrails) in an otherwise fixed Earth-atmosphere system. The ~~cirrus~~ ice water content of the cirrus is adjusted to an optical thickness $\tau = 0.3$ at 550 nm wavelength, and the effective radius of the hexagonal ice particles in this model is set to 20 μm, typical for aged ~~contrail cirrus (Minnis et al., 2013). The~~ contrails (Minnis et al., 2013; Schumann et al., 2017). At TOA, the net instantaneous RF is positive ~~for the LW and "normal" (SW+LW) cirrus cases and~~while the net surface RF_i is negative~~ for SW cirrus~~, consistent with earlier results (see ~~Table 1).~~Figure 2). For comparison, we also consider a 10 % increase in CO_2 (360 to 396 μmol mol^{-1}) ~~again for an otherwise fixed climate system.~~in the same model. Figure 4 and Figure 5 show the instantaneous radiative flux changes and the corresponding heating rates for added SW, LW and ~~normal cirrus and for increased CO_2. Among others, the heating rate profile for cirrus depends strongly on the assumed optical thickness of the cirrus. For thicker cirrus, the LW heating~~

~~increases on average over the cirrus but may get negative at top of the cirrus (Liou, 1986). The large heating rate in the air layer at the fixed-temperature surface reflects the finite net downward radiative fluxes at that surface.~~"normal" (SW+LW) contrail cirrus and for increased CO_2. Figure 8 shows the steady-state temperature response to the radiative disturbance for the three cirrus cases and CO_2.

For contrail cirrus, we see strongly different temperature responses ~~for the~~ in the net, SW and LW versions (Figure 8). The SW ~~and LW cirrus, at least for weak turbulent mixing (Figure 12). The SW cirrus~~contrail causes a slight warming inside the cirrus layer by solar radiation absorption (Stackhouse and Stephens, 1991). The main SW effect ~~of the SW cirrus~~ is a cooling of the lower troposphere culminating at the ~~Earth~~Earth's surface. The LW ~~cirrus~~contrail enhances infrared absorption inside the ~~cirrus~~layer and slightly warms the troposphere below ~~the cirrus~~ by emission from the ~~cirrus~~contrail. In addition, the LW ~~cirrus~~contrail enhances the radiation energy budget at the ~~Earth~~Earth's surface, causing a slight warming, but the SW cooling dominates. ~~Only for strong vertical mixing,~~In the ~~heat induced by~~radiative case, the ~~cirrus~~temperature change is positive in the upper troposphere ~~gets transported downwards quick enough compared to radiative losses to effectively warm~~and negative near the surface. ~~Convective mixing is weak in this example because the cirrus stabilizes the atmosphere below the cirrus.~~ Convective mixing occurs ~~again~~for this atmosphere, with fixed dynamical heating, only in the ~~uppermost~~upper troposphere~~, between the cirrus layer and the TP.~~ where the contrail heating causes unstable stratification. The diffusive mixing distributes the heat nearly uniformly over the troposphere. Without such mixing, the heat induced radiatively by the contrail in the upper troposphere is inefficient in heating the surface.

~~We note that the cirrus also cools the surface in a case with $Q_0 = 0$, i.e. without fixed dynamical heating, for otherwise the same parameters (most important are albedo and SZA), see Figure 5 and Figure 6. In radiative equilibrium without mixing, again, the cirrus warms the tropopause region but cools the lower troposphere and the surface because of dominant SW changes. The given cirrus cools strongest without mixing but cools also with convective adjustment because the cirrus stabilizes the mid troposphere. Only in case of strong and vertically uniform mixing, positive RF causes a positive temperature change throughout the troposphere and at the surface.~~

The CO_2 case shows tropospheric warming as expected (Ramanathan and Coakley, 1978; Manabe and Stouffer, 1980; Ogura et al., 2014). The initial heating, mainly from LW radiation, is positive but small (<0.022 K d^{-1}) in the troposphere and negative in the upper stratosphere with far larger magnitude (-0.6 K d^{-1} at 60 km). ~~The literature shows a range of results for CO_2-induced heating rates (Collins et al., 2006; Dietmüller et al., 2016).~~ Enhanced CO_2 not only heats the troposphere, it also increases the downwelling LW flux reaching the surface. ~~Convective~~For the given atmospheres and disturbances, convective adjustment occurs ~~for this atmosphere~~ only in the middle and in the upper troposphere; the other parts remain stably stratified. The literature shows a range of results for CO_2-induced heating rates (Collins et al., 2006; Dietmüller et al., 2016). The larger global mean upper tropospheric temperature response in climate models (Hansen et al., 1997a) results from amplification by various ~~climate system changes~~feedbacks not included in this model. Global models often show a rather smooth profile of temperature increase in the troposphere, likely because of strong mixing. At high latitudes, reduced vertical mixing under stably stratified conditions, besides sea ice albedo changes, ~~would enhance~~is known to cause enhanced LW warming at the surface from increased CO_2 (Wetherald and Manabe, 1975).

Table 1 lists the~~ computed~~ values for RF_i (at TP, TOA, and surface), RF_a and RF_s at the TP, ΔT_s, and related λ_a, λ_s and efficacy values e_a, e_s, with respect to the CO_2~~, without~~ disturbance, for the radiative and ~~with ~~diffusive ~~mixing. The results for convective mixing are close to those without mixing and not shown, therefore~~cases. The instantaneous and stratospheric adjusted values apply to fixed troposphere and are, hence, independent of tropospheric mixing. The results for the radiative-convective case are close to the radiative case and not shown, therefore.

For CO_2, RF_i is positive throughout the atmosphere. RF_a at the ~~TP~~tropopause is in between the RF_i values at TOA and at the ~~TP~~tropopause, consistent with earlier results (Stuber et al., 2001; Dietmüller et al., 2016). The effective RF_s for fixed climate system is in between the RF_i values at the ~~TP~~tropopause and at the surface.

For contrail cirrus, Table 1 shows that RF_a is small and not much different from RF_i, consistent with Dietmüller et al. (2016). The RF_s values ~~for cirrus ~~differ strongly from RF_a, even with different sign in the radiative case without diffusive mixing. For SW and LW cirrus separately, the ratio $RF_s/RF_{i,TOA}$ increases strongly ~~with vertical~~for diffusive mixing, e.g., from 0.~~22~~23 to 0.~~90~~92 for LW cirrus. At steady state, more and more of the heat induced radiatively by the cirrus reaches the surface and less leaves to space for increased mixing. The temperature sensitivity λ_s is about 40 % smaller with the mixing. Surface heating (or cooling) is more efficient in heating the underlying compartment (larger RF_s/RF_i) than upper tropospheric heating. For the LW+SW ~~cirrus~~contrail, the SW and LW results for RF and temperature add linearly. However, the sensitivities and efficacies change nonlinearly because they are ratios of RF and ΔT_s values. Based on RF_a, the efficacy of SW contrail cirrus is larger than for LW contrail cirrus. Hence, efficacies derived from stratosphere-adjusted RF depend on the heating profiles and on the mixing~~.~~ in the troposphere. Based on RF_s, the efficacies for ~~the well-defined cases ~~SW and LW contrail cirrus are close to unity. They are all close to one, because the ~~cirrus~~cloud and the CO_2 changes considered are small disturbances of the same climate system and the modelled climate systems remain similar also after fast adjustments in ~~all ~~these model cases.

~~Though the nature of the ghost forcing is different, the insight gained in the previous section~~The thin contrail cirrus changes the thermal relaxation properties of the atmosphere only little. It would require a contrail cirrus with optical depth of order one and 100 % cover to cause strong changes of the heat losses to space. Hence, the insight gained for ghost forcing, consistent with Hansen et al. (1997a), helps to understand the temperature changes induced by ~~cirrus.~~radiative heating from contrails. For weak mixing, ΔT_s is highly sensitive to the altitude in which the cirrus heating is induced. Also the dependence of λ on mixing and the usefulness of effective RF_s to estimate ΔT_s with nearly constant λ_s found for ghost forcing, apply similarly for cirrus. Similar efficacies can be expected only for similar atmospheres and ~~strong~~sufficient vertical mixing. ~~A thick added cirrus changes the atmosphere strongly and causes not only additional warming but also reduces heat loss from the surface and from the atmosphere below the cirrus to space. ~~In all cases, we find that the effective RF_s is in between ~~the~~

~~values of~~ RF_i at the ~~TP~~tropopause and at the surface. This finding may be helpful for estimating RF_s for given instantaneous RF.

~~In the contrail climate study with a global circulation model by Ponater et al. (2006)~~Different from ghost forcing, the contrails change the optical properties not only in the infrared but also in the solar range. The effects of the contrails are, of course, sensitive to surface albedo and SZA which were irrelevant for ghost forcing. However the model behaves still quasi linearly. An increase of contrail coverage, e.g., from 3 to 12 % changes the efficacies by 0.38%, 3.87% and -2.76% for SW, LW and SW+LW contrails in the radiative case and by smaller values for the two cases with mixing. Some of the cases were recomputed with DISTUF instead of the more efficient two-stream solver. For CO_2 and SW+LW contrail cirrus, the values in Table 1 differ by <8 % in magnitude between the two solvers.

Ponater et al. (2005) studied contrail climate sensitivity with a global circulation model. They show a plot of the zonal mean vertical cross-section of annual mean temperature response in the equilibrium climate ~~shows that the contrail-induced warming is a maximum in the upper troposphere and limited to the latitude band in which contrails formed. Hence, the mixing was not strong enough to disperse the contrail-induced warming uniformly over the troposphere.~~which shows that the contrail-induced warming is largest in the upper troposphere and limited to the latitude band in which contrails formed. The maximum in the upper troposphere cannot be explained by local release of latent heat, because the amount of water condensing during cloud formation at those low-temperature levels is small. The pattern with enhanced temperature for contrails is more pronounced than for a similar CO_2 disturbance simulation (see Figure 1 of Ponater et al. (2006b)). Hence, the mixing was likely not strong enough to disperse the contrail-induced radiative heating uniformly over the troposphere. The different efficacies found by Rap et al. (2010a) and by ~~Ponater et al. (2006) may be caused by different ratios of SW to LW RF magnitudes and different vertical mixing in the different models, besides~~Ponater et al. (2005) may be caused by different vertical mixing in the different models, besides different ratios of SW to LW RF magnitudes and different feedbacks.

Figure 9 illustrates the ~~altitude, scale and mixing dependent~~ timescales of temperature relaxation inside the atmosphere for a non-steady case. Here we show temperature profiles as a function of time starting from steady state for the given contrail cirrus ~~and given mixing model~~ over an adiabatic ~~land~~ surface, e.g. over land, after the ~~cirrus~~contrail is suddenly taken away. The times needed to reach half the initial values, derived from plots of the results versus time, are 0.8 d, 8 d, and 50 d for the temperature at the surface, on average in the contrail layer, and in the troposphere, respectively, for the radiative case. The mean tropospheric halving time is 12 d for the radiative-diffusive case. As expected from the ghost forcing results, the temperature change returns to zero most rapid at the surface ~~(reaching half its initial value within one time step, 0.25 d);~~; the temperature within the ~~cirrus~~contrail layer also returns to zero quickly ~~(6.5 d)~~ because of the relatively small geometrical ~~cirrus~~contrail layer depth, while the ~~thicker~~ troposphere in whole needs ~~22.5 d~~nearly 2 months to reach half its initial mean value. The larger troposphere value~~.~~ is a consequence of its larger thickness. For constant surface temperature~~, the relaxation times would be smaller. Convective mixing does not change the results much for this atmosphere. The~~ and the diffusive mixing~~ reduces both the temperature maximum and the mixing relaxation times scales for local temperature disturbances considerably~~, the troposphere reaches half its initial value after 2.7 d. Of course, thermal inertia of an ocean would increase heat residence times to many years (Hansen et al., 1985). The example illustrates quick losses of energy by radiation to space, which gets enhanced by mixing within the troposphere.

4 Summary, Implications and ~~discussions on regional effects and~~ Conclusions

The surface temperature sensitivity to small climate disturbances has been investigated in an idealized climate model without feedbacks except by temperature changes and lapse-rate dependent convection. The model is a one-dimensional representation of the mid-latitude summer atmosphere with constant insolation. Fixed dynamical heating is imposed to study small disturbances of an undisturbed atmosphere in steady state. The boundary conditions prescribe either fixed surface temperature or zero heat flux through the surface. The fixed-surface-temperature case is used to simulate fast adjustment processes; it provides the effective RF estimate. The zero-heat-flux case simulates an atmosphere in thermal equilibrium with the compartment below the surface. Disturbances considered are layer heating (“ghost forcing”), a prescribed contrail cirrus layer, and a 10 % increase of CO_2 mixing ratio. Radiative fluxes are computed with an efficient two-stream solver from libRadtran. Diffusive fluxes are driven by the potential temperature gradient. The diffusivity is set either constant or lapse-rate dependent to simulate vertical diffusive mixing, e.g., from large scale eddies, or convective adjustment in unstable layers of the atmosphere. The model response is quasi linear in the magnitude of the disturbances for fixed mixing properties but nonlinear with lapse-rate dependent convective adjustment. From the model results, the ratio of layer-mean temperature changes to heating rates is used to characterize the time scales for radiative relaxation. Model results for various boundary conditions are used to compute instantaneous, stratosphere-adjusted, and effective RF, i.e., RF_i, RF_a, and RF_s.

~~The results have obvious implications. If we assume forcing by a regional cirrus change and advection by horizontal wind, then any surface cooling or warming will be limited regionally to the immediate neighborhood of the domain with cirrus changes while the upper troposphere warming may travel over large distances. The radiative forcing by cirrus contributes to long-term global warming only when the heat captured by the cirrus reaches the ocean. A globally uniform heating from localized forcing is unlikely unless advection and mixing occur at timescales far shorter than radiative relaxation. Advection of heat from cirrus or contrail warming has been noted in previous simulations~~The model results provide thermal relaxation time scales of the order of hours near the surface, of about one to two weeks in the upper troposphere and of order a month in the lower stratosphere. After fast adjustment, RF_s is nonzero and smaller in magnitude than RF_i. Final thermal equilibrium with an ocean below the surface would be reached far later, after many years to centuries. This final state is simulated with the zero-flux boundary condition. The ratio $RF_s/RF_{i,TOA}$ measure the fraction of the instantaneous energy flux change at TOA available after fast adjustments for long term heating of the compartment below the surface. The ratio RF_s/RF_i depends on the height of where the disturbance induces radiative heating. For zero turbulent mixing, RF_s/RF_i decreases from large values (80 % in the case simulated) for heating directly at the surface to small values (<5 %) above the tropopause. For the diffusive vertical mixing in the troposphere, the ratio RF_s/RF_i approaches a constant of order 60 %. Hence, a large fraction (about 40 %) of the initial energy flux disturbance radiates to space and cannot heat the compartment below the surface. The temperature sensitivity varies with layer height if defined relative to RF_i but is constant relative to RF_s. RF_s controls the transient heating rate of the compartment below the surface.

The contrail layer introduces a positive instantaneous RF at TOA in the LW and a negative RF in the SW range with a small positive net RF. The heating rate profiles in the LW and SW ranges are different with larger magnitude near the surface for SW than for LW flux changes. At the surface, the net RF is negative. As a consequence, the temperature sensitivities differ between the LW and SW forcing parts and change depending on the degree of mixing. For zero mixing, the surface energy budget with SW cooling may dominate the LW warming at the surface. Hence, the temperature sensitivity could be negative.

Taking the temperature sensitivity to the CO_2 disturbance in the same atmosphere for reference, we find that the contrail efficacies based on RF_i differ strongly from unity while those based on RF_s are near one. The temperature sensitivity for the sum of SW and LW RF is the ratio of differences of large opposing values and hence, sensitive to minor system changes. So, for contrails with large and nearly cancelling LW and SW effects, no simple relationship between radiative forcing and temperature change may exist. The findings may apply also for other disturbances.

It is important to keep in mind that the results presented in this study are from a conceptual model. The results are model and case dependent. The fast adjustment and even more the equilibrium responses in general depend on many feedbacks to the temperature-mediated climate system changes, not included in this study. Any change in the model setup, the reference atmosphere or the nature and magnitude of the disturbances would change the results at least quantitatively. Hence, though our study shows the principle importance of mixing for climate sensitivity to contrails, we cannot say how important mixing is for real world cases quantitatively. Ultimately, this requires careful simulations with a comprehensive climate model.

One climate model study (Ponater et al., 2005) indicates that the mixing of contrail-induced warming may be indeed weak and insufficient to mix the heat over the troposphere uniformly. Differences between the efficacy estimates from various studies may partly be caused by different mixing rates in the models used. Future studies should document the mean radiative, adjective and turbulent energy fluxes, including the TOA and surface energy budgets, to allow for analysis of the relative importance of various energy transport mechanisms for climate sensitivity. It may be of interest whether a correlation between contrails and atmospheric mixing conditions exist. Contrail-induced heating in the upper troposphere during calm weather may contribute less to surface warming than the same forcing in a strongly mixing weather situation. Shorter time scales of SW-induced temperature changes near the surface may lead to a dominance of SW surface cooling relative to LW warming regionally where contrails form, while remaining LW warming may dominate after advection downstream at larger distances. The importance of advection of heat induced by contrail warming has been noted previously (Ponater et al., 1996; Rind et al., 2000), ~~but the role of radiative cooling has not yet been discussed. Spatial variability in the forcing/response relationship has been derived from climate models for aerosol forcing~~and also for other disturbances

(Shindell et al., 2010)~~. Hence, efficacy differences are to be expected on where over continents and oceans~~, but the ~~cirrus formed. For small mixing and radiation relaxation time scales also the time scales of the disturbances itself (e.g., minutes to days~~potential for ~~contrails and cirrus) influence the mean efficacy of the related RF, because local warming radiates more quickly to space than well mixed warming.~~

~~The results presented so far were obtained including fast temperature changes and mixing for otherwise fixed atmosphere, without taking other changes of the climate system (feedbacks) into account. As a consequence of temperature change, the climate system will change in many respect (Stephens, 2005). Here we add some discussion to this. Because of~~ different ~~heating profiles and incomplete mixing, temperature change profiles are different and, hence, feedbacks for cirrus will be different from those for CO_2.~~

~~For illustration, we apply our model also with absolute humidity adapted to temperature changes for fixed relative humidity (Manabe and Wetherald, 1967). For cirrus, because of local warming, such a change enhances humidity mainly in the cirrus itself. Water vapor is a particularly efficient greenhouse gas near the TP, and added water vapor increases the surface temperature (Shine and Sinha, 1991), consistent with our results, see Figure 14. RF_s is computed for the atmosphere with fixed humidity assuming that the change in humidity (e.g., because of ocean warming) is a slow process. Table 2 lists the temperature changes, climate sensitivities and efficacies in steady state with and without humidity feedback and a feedback factor F, i.e., the ratio of ΔT_s with and without humidity changes.~~

~~The efficacies and feedback factors for cirrus with LW warming and SW~~ radiative warming and cooling ~~heating rates are highly sensitive to small system changes ("ill-conditioned") because the RF is the difference of two large contributions and the sign of ΔT_s and RF may differ when both are close to zero. We see that the efficacies and feedback factors for SW and LW cirrus differ from one. In contrast to efficacies for RF_a, the efficacies for RF_s in the atmosphere with humidity changes are larger for LW cirrus than for SW cirrus. Both are different from one. Hence, neither RF_a nor RF_s are direct measures of the equilibrium surface temperature change. In the cirrus case, LW forcing gets enhanced while SW forcing gets reduced by climate system changes from changed humidity. Kashimura et al. (2017) investigate surface cooling by added stratospheric aerosol and also find reduced SW RF by reduced humidity and low-level clouds. Ultimately, the role of climate system changes for the RF cannot be determined with a simple model. It requires simulations with a comprehensive climate model.~~

~~5 Conclusions~~

~~Surface temperature changes induced by radiative disturbances depend on the vertical distribution of the radiative heating induced by the disturbances in the troposphere. Since cirrus introduces warming and cooling contributions~~effects at different altitudes~~, the surface temperature response to radiative forcing by added cirrus and contrails is particularly sensitive to the vertical heating rate profile. It requires strong vertical heat transport by mixing to distribute the induced heat uniformly over the whole troposphere. The mixing~~ has ~~to act at time scales quicker than the radiative heat transfer to avoid loss of energy by~~

~~radiation to space before the heat can reach the surface. Cirrus tends to stabilize the atmosphere with reduced convective mixing, enhancing the sensitivity to the vertical distribution of the radiative heating.~~

~~This paper~~not yet been discussed~~ the relationship between radiative forcings and surface temperature changes in a qualitative manner based on a radiative-convective-diffusive model. Various RF versions are considered, including instantaneous, stratosphere adjusted, and effective RF, i.e., RF_i, RF_a, and RF_s. Here, RF_s is computed for fixed surface temperature and the limited set of adjustments represented in the model. After adjustment by thermal relaxation, the RF_s was found to be in between the RF_i values at TOA and at the surface and smaller in magnitude than the corresponding RF_a values. As an extreme, for weak tropospheric mixing, added cirrus may cool the surface even when RF_i and RF_a suggest warming. In agreement with earlier studies, we find that the climate sensitivity to RF_a varies strongly between the various forcing types while the sensitivity to RF_s is closer to constant. However, when the climate system changes beyond what is included in the fast adjustments considered for RF_s, e.g., by humidity changes during ocean warming, the efficacies vary between the forcing types also for RF_s. For cirrus including LW and SW effects, no simple relationship between net radiative forcing and temperature change exists.~~

~~The radiative relaxation time scales of the disturbance-induced temperature profile changes are of order hours near the surface to months in the mid stratosphere. Hence, temperature changes induced by cirrus near the surface are short lasting and may be more regionally limited, while upper tropospheric temperature changes last longer and may spread over a larger part of the Earth.~~

~~The classical RF concept assumes sufficiently strong mixing within the troposphere, i.e., mixing time scales shorter than the time scales of thermal relaxation by radiation. One climate model study (Ponater et al., 2005) indicates that the mixing of contrail-induced warming is too weak to mix the heat over the troposphere uniformly. Hence, the contrail warming is distributed over a smaller domain and lasts shorter than for CO_2 and this, besides different feedbacks, may cause different efficacies~~.

These findings may have implications for the assessment of the climate impact of aviation by ~~contrail cirrus.~~contrails. So far, equilibrium warming from contrails is computed using estimates of RF (RF_i or RF_a) together with a CO_2 climate sensitivity corrected by a contrail efficacy (Ponater et al., ~~2006~~2006a; Lee et al., 2009a; Frömming et al., 2012). ~~The net RF for cirrus is often far smaller than the magnitude of its SW and LW parts. In this~~Our study ~~we found~~suggests that the ~~efficacies for SW and LW parts may differ. Hence, the~~ efficacy~~-weighted RF may~~ should be ~~much~~ different ~~from previous estimates~~for SW and LW forcings. This may be important for comparison of the climate impact of different contrail cases, e.g., for different diurnal traffic cycles or different route settings.

This study ~~adds~~provides further insight into ~~why~~known limitations of the RF model ~~is not a universally applicable method to estimate and compare the climate change contributions from various disturbances.~~approach. Hence, better approaches are needed. A suggestion for an alternative to the RF concept, based on a new temperature forcing concept, will be described in a follow-on paper (submitted) to this study.

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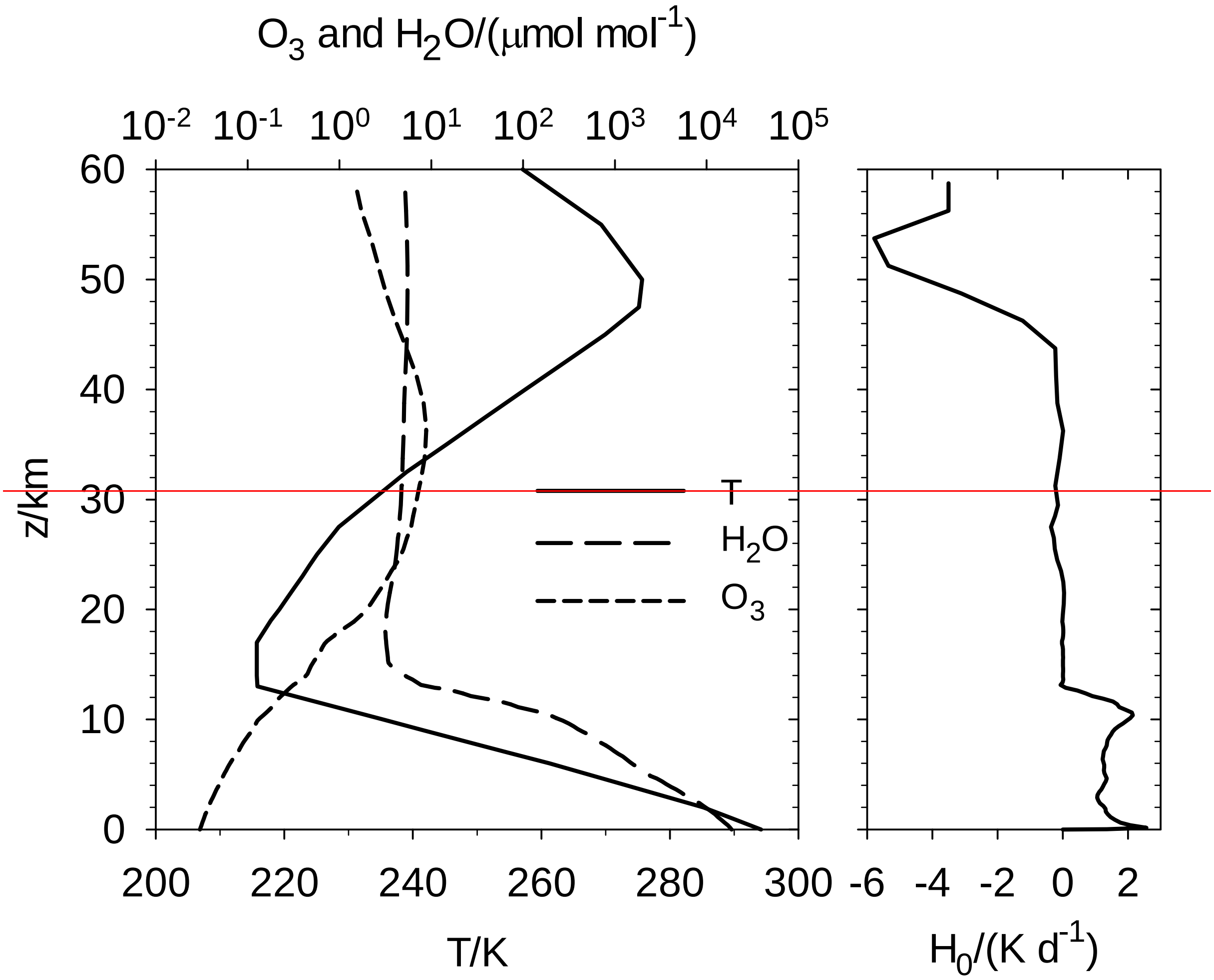
Table 1. Radiative Forcing (RF~~) for cirrus and CO_2 for fixed climate system (i: instantaneous at tropopause (TP), top of atmosphere (TOA), and surface (SUR); a: adjusted at TP; s: effective at TP), equilibrium air)*)~~, surface temperature ~~changes ΔT_s, (assuming instantaneous heat mixing between surface and lowest model layer) and~~change ΔT_s, sensitivity parameters λ and efficacies e relative to adjusted and effective RF_a and RF_s, for contrail cirrus and CO_2 for the given model. The first four rows are the radiative cases with zero turbulent fluxes, the last four rows apply for ~~strongly~~ the radiative-diffusive cases. The instantaneous and adjusted RF values are the same for both ~~mixing versions.~~cases. Negative λ and e values~~ for cirrus~~ are considered ill-conditioned because highly sensitive to small changes in forcing and mixing contributions.

	RF_i W m^{-2}	$RF_{i,TOA}$ W m^{-2}	$RF_{i,SUR}$ W m^{-2}	RF_a W m^{-2}	RF_s W m^{-2}	ΔT_s K	λ_a K W^{-1} m^2	λ_s K W^{-1} m^2	e_a 1	e_s 1	$RF_s/RF_{i,TOA}$ 1
								radiative case			
CO_2	0.83	0.41	0.07	0.72	0.~~27~~26	0.12	0.~~17~~16	0.45	1.~~00~~	1.~~00~~	0.~~37~~64
SW Cirrus	-0.~~81~~49	-0.~~80~~48	-0.~~56~~46	-0.~~81~~49	-0.~~63~~48	-0.~~28~~22	0.~~35~~44	0.~~45~~44	2.~~12~~69	0.99	~~0.79~~1.00
LW Cirrus	0.92	0.88	0.09	0.90	0.20	0.09	0.10	0.45	0.60	0.99	0.~~22~~23
Cirrus	0.~~11~~43	0.~~08~~40	-0.~~47~~37	0.~~10~~42	-0.~~43~~28	-0.~~19~~13	~~-2.00~~0.30	0.~~45~~44	~~-12.09~~1.83	0.~~99~~98	~~-4.49~~0.70
							radiative~~ and ~~-diffusive case				
CO_2	0.83	0.41	0.07	0.72	0.70	0.19	0.26	0.~~41~~26	1.~~00~~	1.~~00~~	~~0.97~~1.70
SW Cirrus	-0.~~81~~49	-0.~~80~~48	-0.~~56~~46	-0.~~81~~49	-0.~~80~~49	-0.~~21~~13	0.~~26~~27	0.~~30~~26	1.~~02~~04	1.00	~~0.99~~1.03
LW Cirrus	0.92	0.88	0.09	0.90	0.81	0.21	0.24	0.~~40~~26	0.92	1.00	0.~~90~~92
Cirrus	0.~~11~~43	0.~~08~~40	-0.~~47~~37	0.~~10~~42	0.~~01~~32	0.~~00~~08	0.~~04~~20	~~-~~0.~~02~~26	0.~~15~~79	1.00	0.~~15~~79

~~**Table 2.** RF_s in the atmosphere with fixed humidity and temperature changes ΔT_s without and with humidity feedback (first 4 and last 5 columns), for radiative and for radiative-diffusive equilibrium (first and last 4 rows). For both feedback variants, the table lists: ΔT_s, λ_s and e_s (symbols as in Table 1); the last column is the feedback factor F, i.e., the ratio of ΔT_s with and without humidity changes. The efficacies and feedback factors for cirrus including LW and SW effects are again considered ill-conditioned.~~

	~~fixed H_2O~~				~~fixed RH~~			
	~~RF_s~~	~~ΔT_s~~	~~λ_s~~	~~e_s~~	~~ΔT_s~~	~~λ_s~~	~~e_s~~	~~F~~
	~~$W\ m^{-2}$~~	~~K~~	~~$K\ W^{-1}\ m^2$~~		~~K~~	~~$K\ W^{-1}\ m^2$~~	~~1~~	~~1~~
~~radiative~~								
~~CO_2~~	~~0.27~~	~~0.12~~	~~0.45~~	~~1.00~~	~~0.45~~	~~1.71~~	~~1.00~~	~~3.80~~
~~SW Cirrus~~	~~-0.63~~	~~-0.28~~	~~0.45~~	~~0.99~~	~~-0.57~~	~~0.90~~	~~0.52~~	~~2.02~~
~~LW Cirrus~~	~~0.20~~	~~0.09~~	~~0.45~~	~~0.99~~	~~0.53~~	~~2.66~~	~~1.55~~	~~5.95~~
~~Cirrus~~	~~-0.43~~	~~-0.19~~	~~0.45~~	~~0.99~~	~~-0.05~~	~~0.11~~	~~0.06~~	~~0.24~~
~~radiative-diffusive~~								
~~CO_2~~	~~0.70~~	~~0.19~~	~~0.26~~	~~1.00~~	~~0.34~~	~~0.49~~	~~1.00~~	~~1.85~~
~~SW Cirrus~~	~~-0.80~~	~~-0.21~~	~~0.26~~	~~1.00~~	~~-0.41~~	~~0.51~~	~~1.04~~	~~1.93~~
~~LW Cirrus~~	~~0.81~~	~~0.21~~	~~0.26~~	~~1.00~~	~~0.45~~	~~0.55~~	~~1.13~~	~~2.10~~
~~Cirrus~~	~~0.01~~	~~0.00~~	~~0.24~~	~~0.92~~	~~0.04~~	~~2.61~~	~~5.35~~	~~10.70~~



*) Index i for instantaneous values at tropopause (TP), top of atmosphere (TOA), and surface (SUR); a for adjusted at TP; and s for effective at TP.

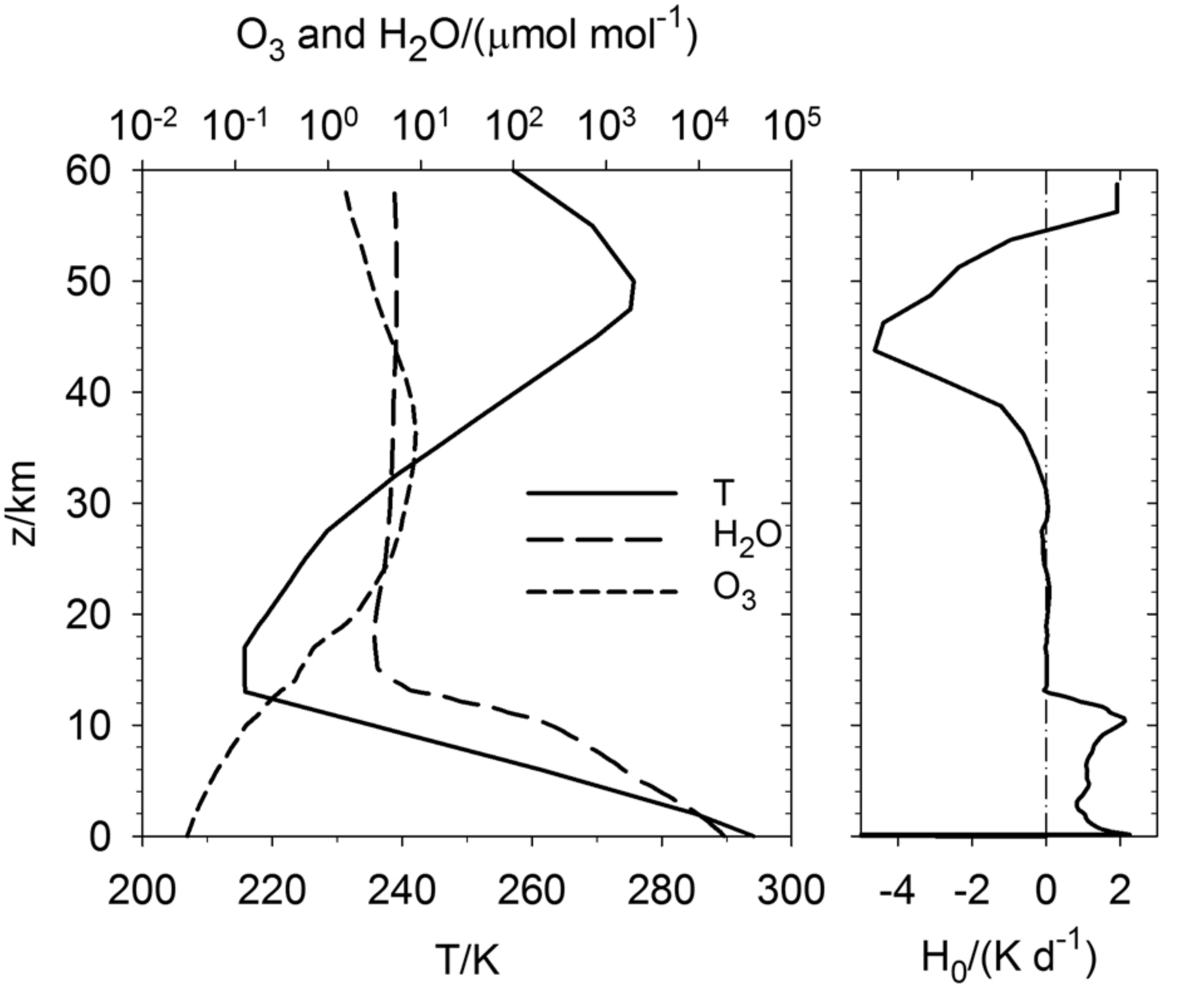


Figure 1. Temperature T of the mid-latitude summer standard atmosphere versus height z, together with water vapor and ozone molar mixing ratio (O_2: 0.2002 mol mol^{-1}; CO_2: 360 µmol mol^{-1}), and heating rate $H_0 = Q_0/(\rho\ c_p)$ keeping the atmosphere at steady-state~~, for fixed~~. Because of TOA radiative imbalance, H_0 is strongly negative at the surface ~~temperature, albedo 0.3, and cos(SZA)=0.25. In the~~; in the mass-weighted average, H_0 = -1.~~39~~25 K d^{-1} in the troposphere and -0.~~062~~057 K d^{-1} in the stratosphere, balancing the summer warming.

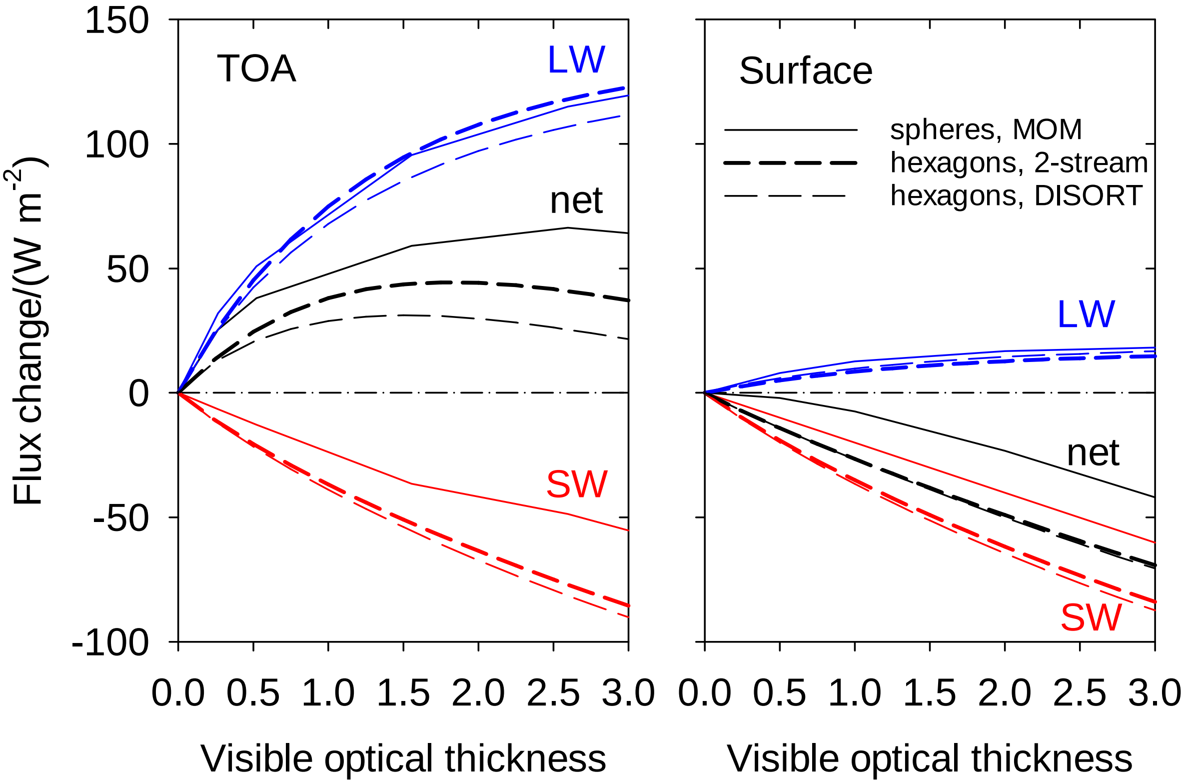
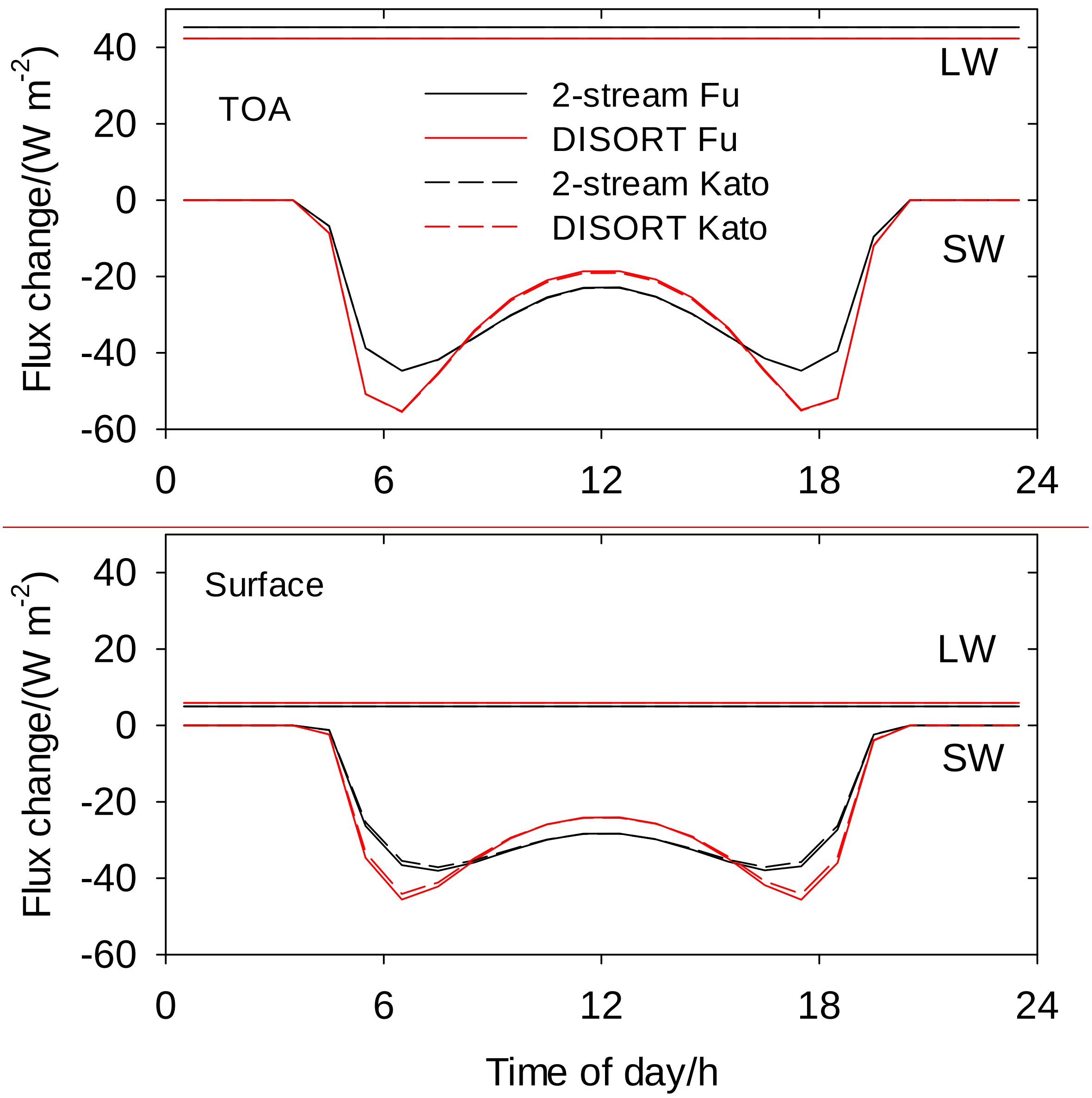


Figure 2. Day-mean ~~flux changes versus 550-nm optical thickness τ for~~cloud radiative effects from a homogeneous ~~cirrus~~contrail layer at 10 to 11 km altitude versus 550-nm optical thickness. The contrail cirrus is assumed to be composed of spheres (Meerkötter et al., 1999) or hexagons (Fu and Liou, 1993)~~,~~. The cloud radiative effect is the flux difference relative to the cloud-free atmosphere, and computed with matrix operator method (MOM; ~~(Plass et al., 1973)~~Plass et al. (1973)), two-stream, and discrete ordinate (DISORT) solvers and the Fu & Liou parametrization for molecular absorption, for daily mean at 45°N, 21 June, standard mid-latitude summer atmosphere, over a surface with albedo 0.2, and fixed surface temperature ~~equal to the surface atmosphere temperature~~ (294.2 K). Differences between the SW (LW) fluxes ~~for these~~from the two-stream and DISORT solvers are ~~of order 10 to 20 %, but DISORT takes orders of magnitude more computing time.~~< 6% (<19%).


TOA
LW
SW
2-stream Fu
DISORT Fu
2-stream Kato
DISORT Kato
Surface
Flux change/(W m-2)
Time of day/h
40
20
0
-20
-40
-60
0
6
12
18
24

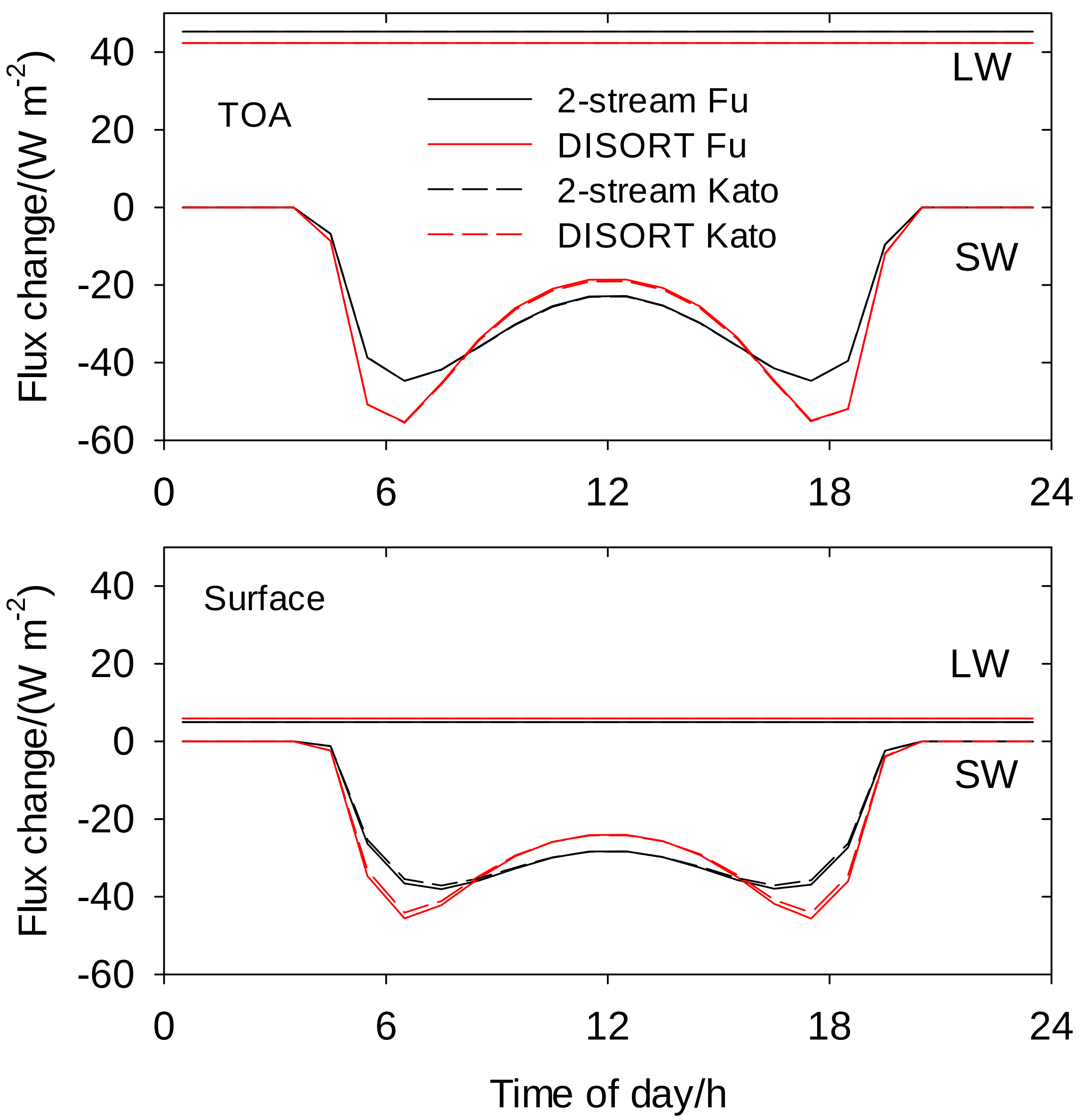
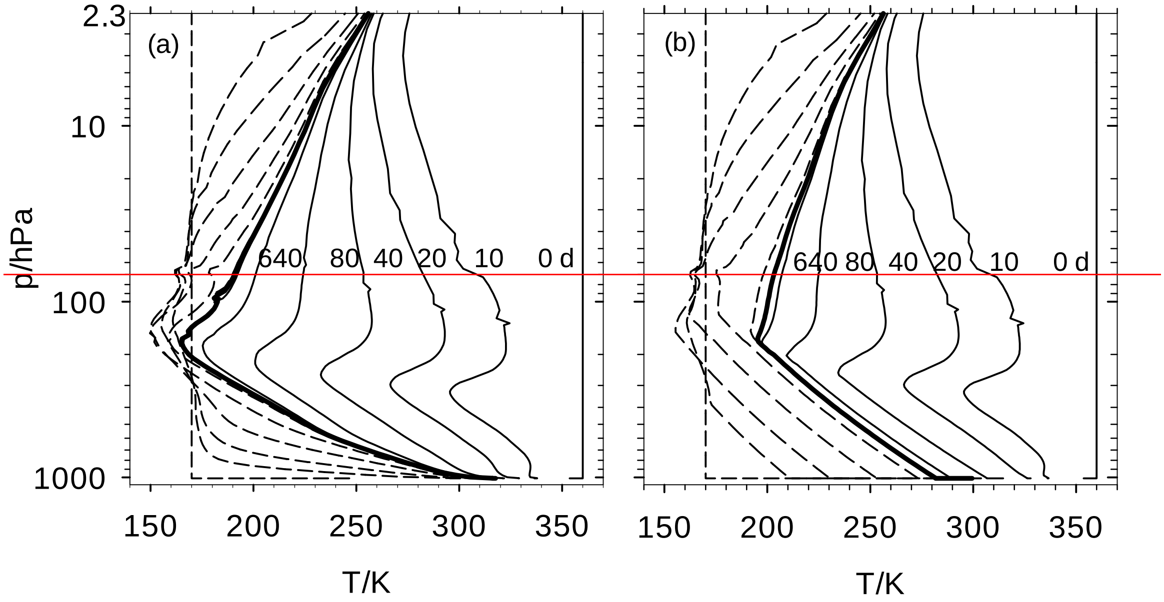
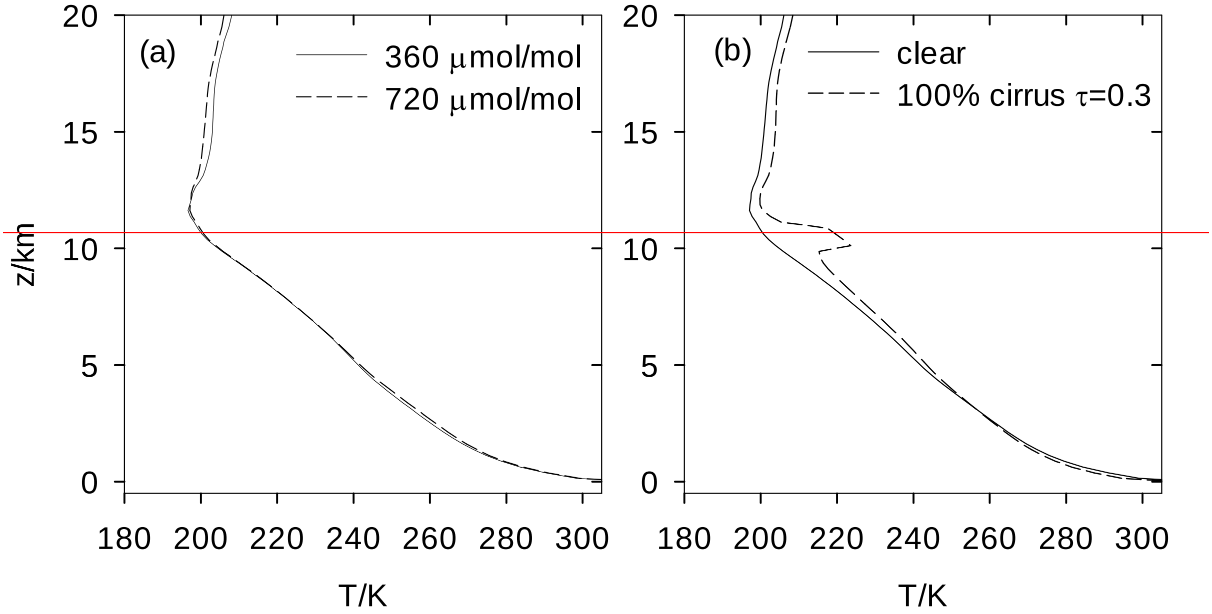


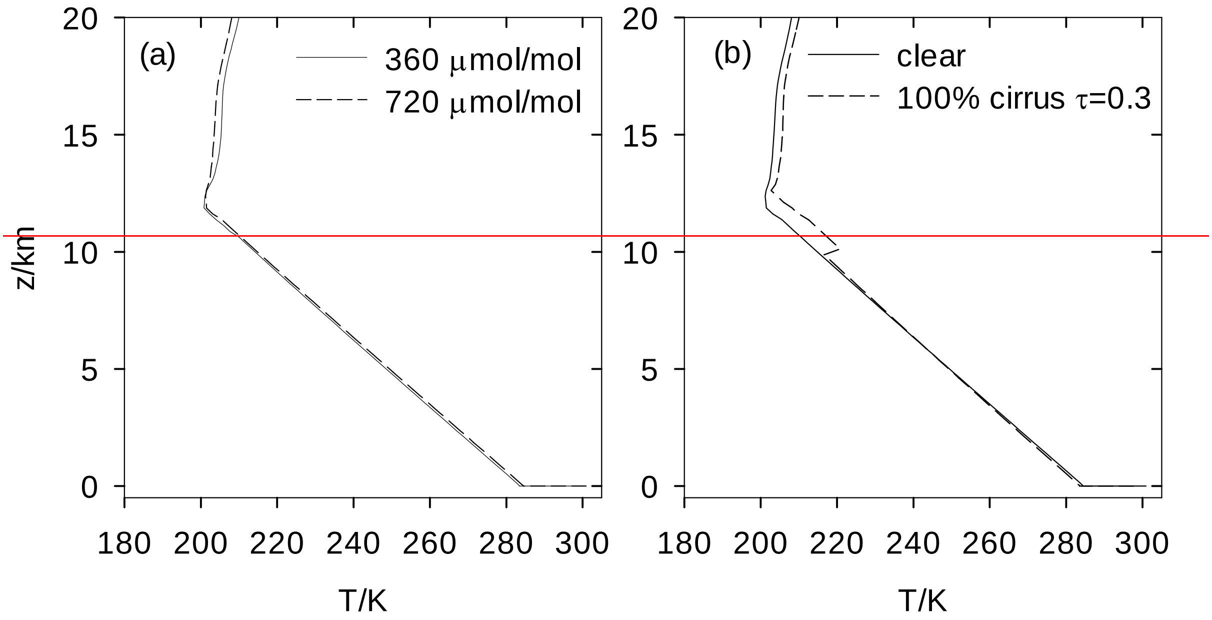
Figure 3. LW and SW flux changes versus time of day at TOA and at the surface, for two-stream and DISORT solvers, and for Fu & Liou and Kato shortwave molecular absorption parametrizations. The model parameters are the same as in Figure 2, ~~for~~ τ =0.5. The flux differences for different molecular absorption models of Fu and Kato are far smaller than between the two-stream solver and DISORT.



~~**Figure 4.** Temperature profiles versus pressure altitude (about 0 to 40 km height) starting from 170 K (dashed) and 360 K (full curves) initially, for comparison with Manabe and Strickler (1964), showing the approach to radiative equilibrium, (a) for pure radiative equilibrium and (b) with convective mixing. The model is applied for the cloud-free and aerosol-free mid-latitude summer atmosphere composition, with tropospheric CO_2 mixing ratio set to 360 µmol mol^{-1}, cos(SZA) = 0.25, Lambertian surface with albedo = 0.3 and emissivity= 1. Curves are shown for times 0, 10, 20, 40, ..., 640 d as partially identified by labels. The thick curves show the temperatures after 640 d. The final temperatures from the two initial conditions differ by less than 10^{-3} K in the troposphere and by 0.2 K near 100 hPa.~~



~~**Figure 5**. Pure radiative equilibrium temperature profiles versus height (a) for reference and for doubled CO_2 mixing ratio. (b) Same for reference atmosphere and atmosphere with a 100 % coverage by a cirrus layer at 10-11 km height with 550 nm optical thickness of 0.3. The doubled CO_2 causes strong stratospheric cooling and a weak tropospheric warming. The cirrus causes a warming in the stratosphere and upper troposphere but a cooling in the lower troposphere and at the surface.~~



~~**Figure 6**. Same as Figure 5 with convective mixing. The warming/cooling effects have still the same signs. Convection causes heat exchange leading to warming in the mid-troposphere. With convection, a temperature inversion forms below the given cirrus layer.~~

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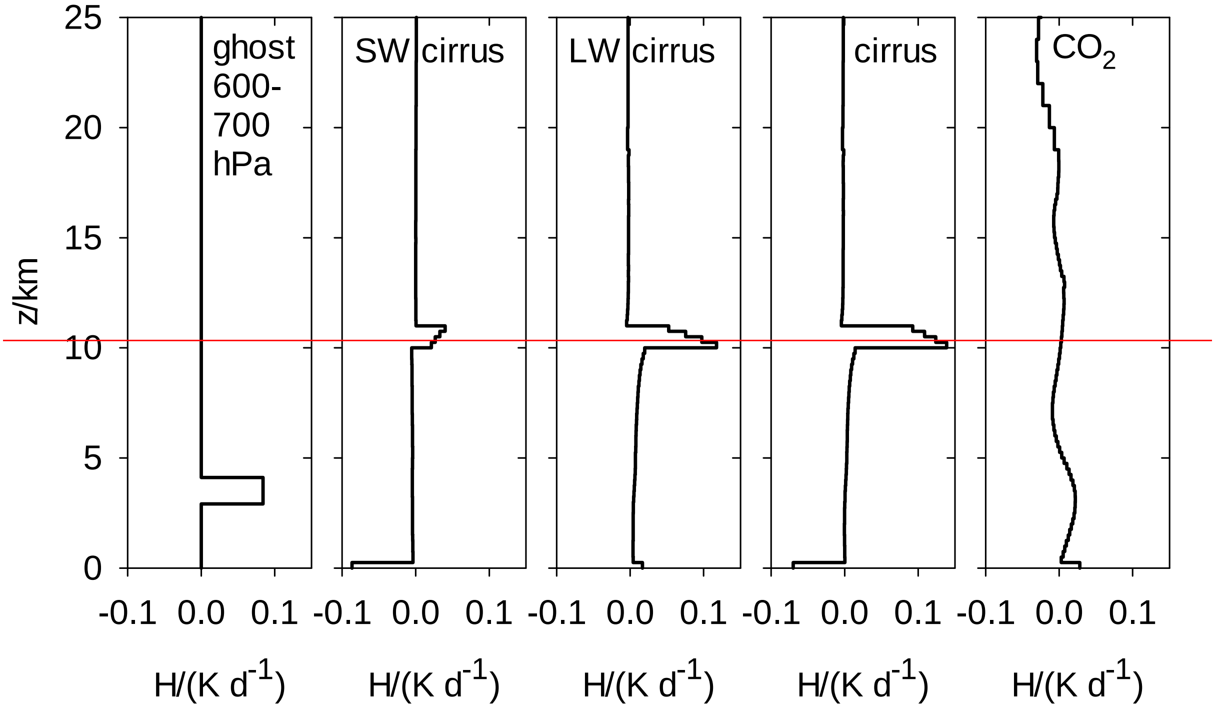


~~Figure 7: Initial radiative heating rates H(t=0, z) versus height z for a ghost forcing example, for SW cirrus, LW cirrus, normal cirrus, and for a CO_2 disturbance.~~

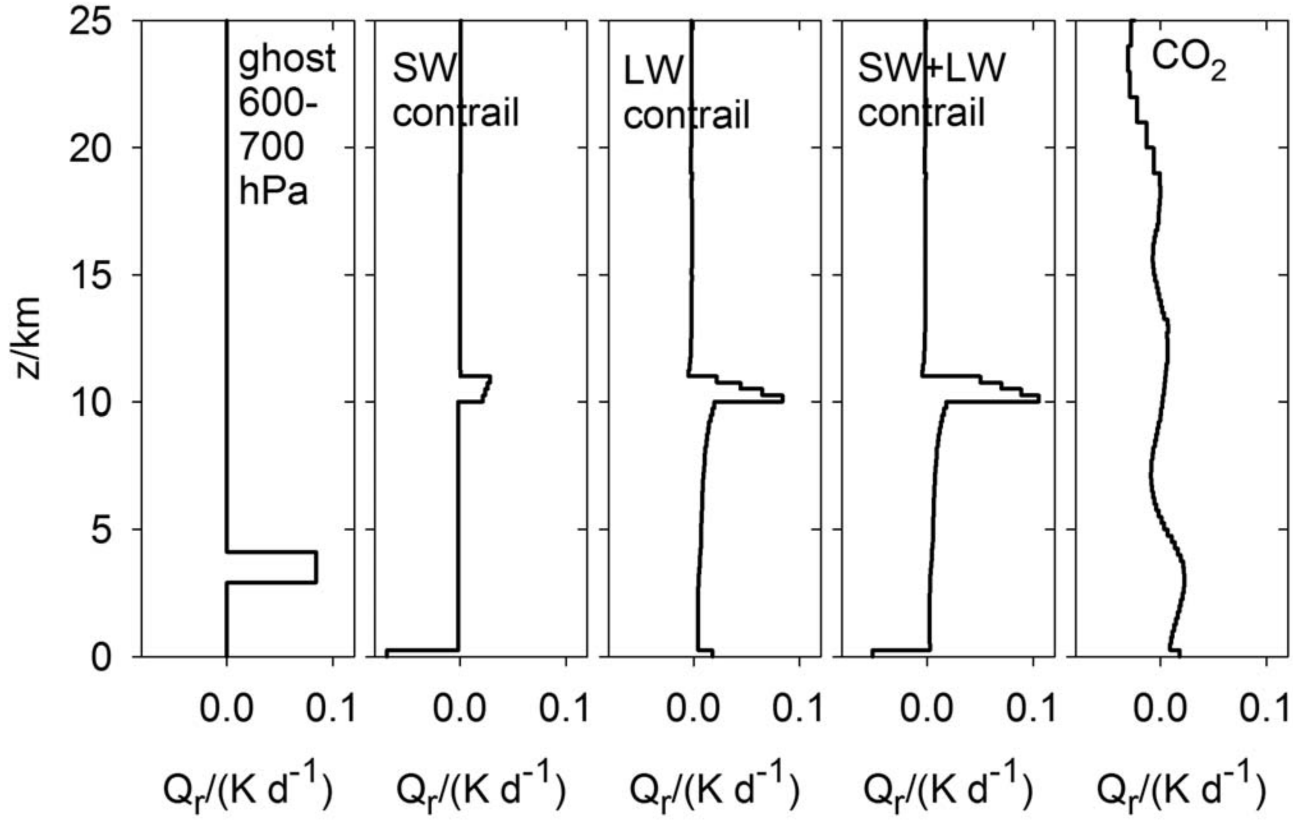
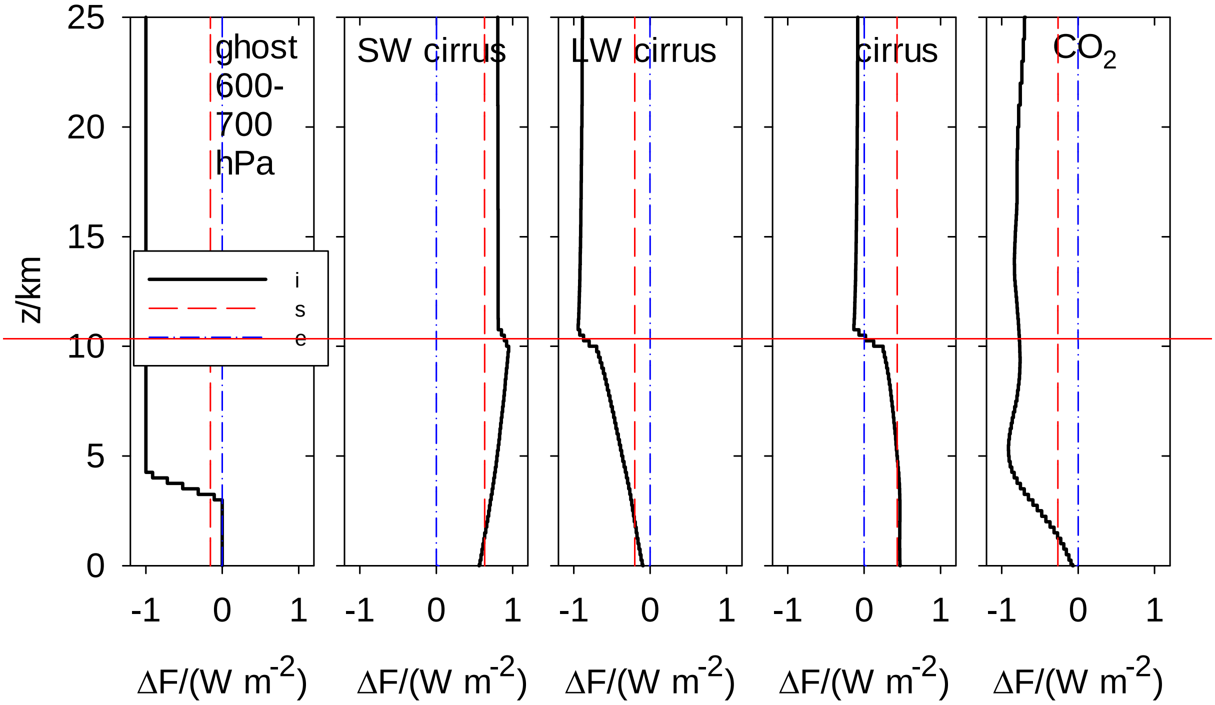


Figure 4: Initial radiative heating rates H(t=0, z) versus height z for a ghost forcing example, for SW, LW and SW+LW contrail cirrus, and for a CO_2 disturbance. For plotting, the local heating rate induced by the nonzero radiative fluxes at the fixed-temperature surface is distributed over the lowest 275 m height (same heat capacity as 1 km thick cirrus layer at lower pressure).



ghost
600-
700
hPa
SW cirrus
LW cirrus
cirrus
CO2
i
s
e
z/km
25
20
15
10
5
0
-1
0
1
ΔF/(W m-2)

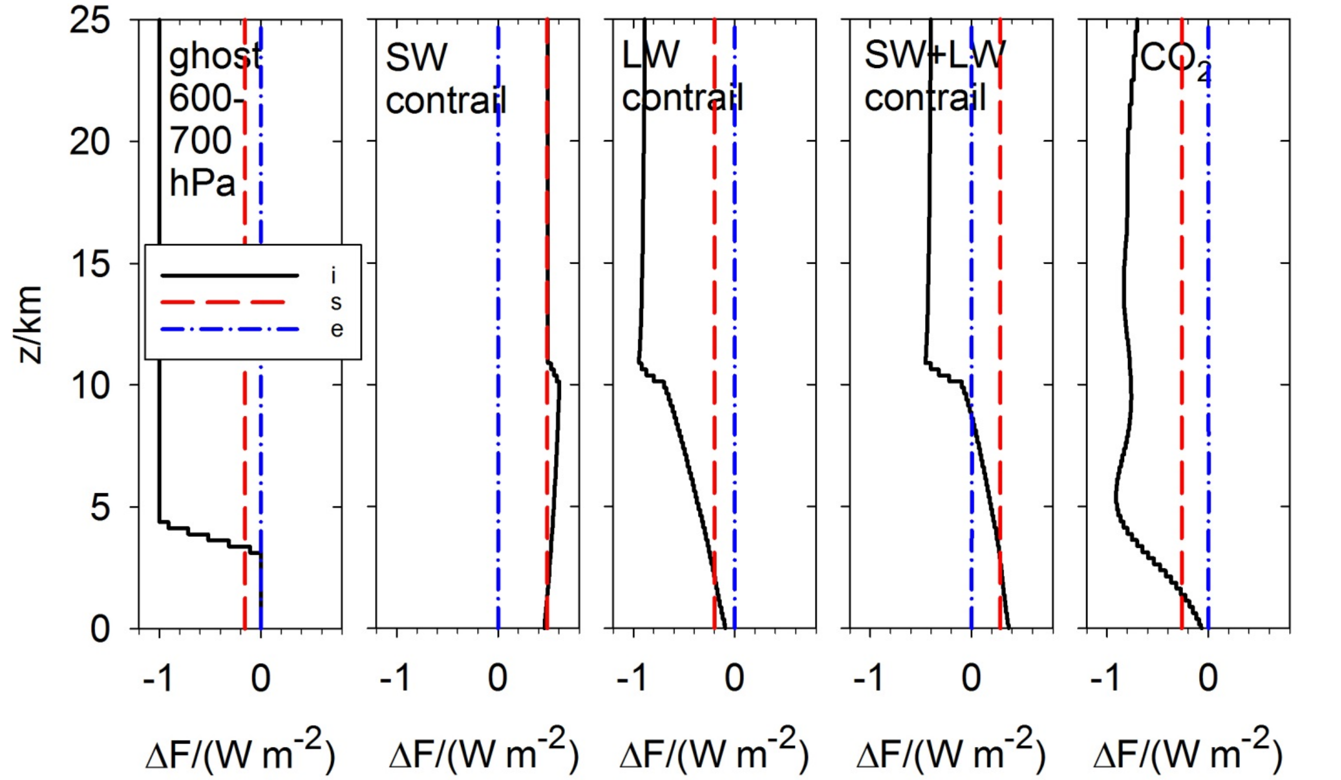
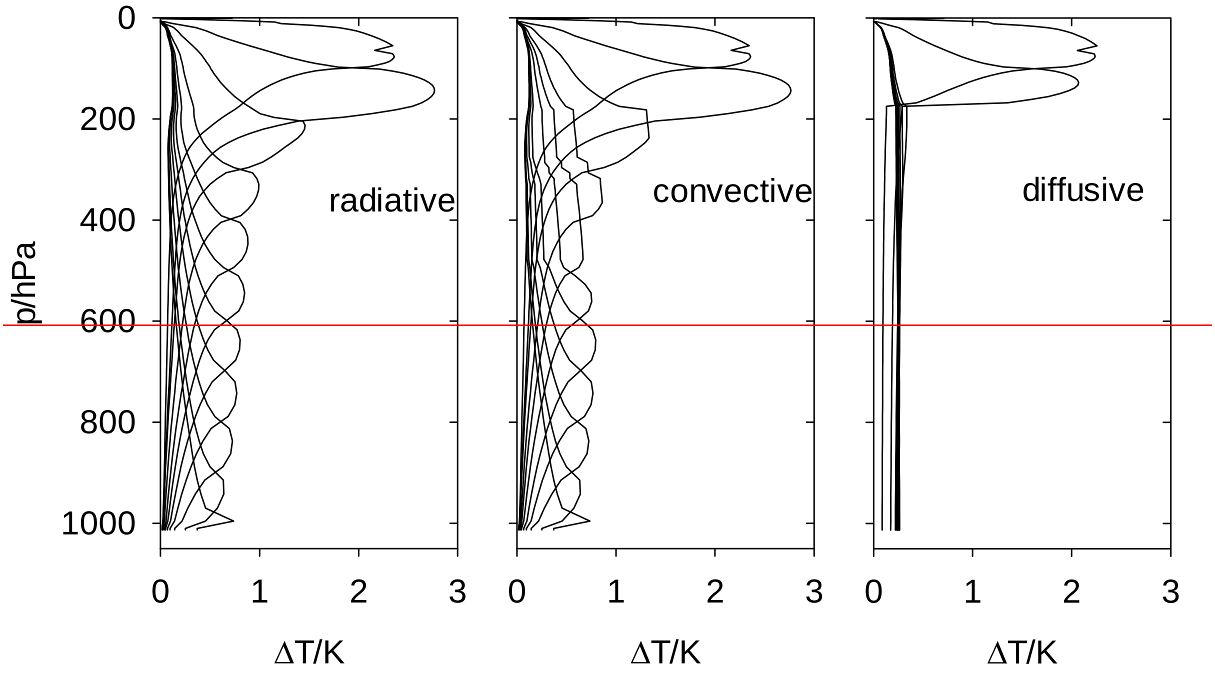


Figure 5. Initial ~~(instantaneous)~~ and final ~~(stratosphere adjusted or equilibrium) net~~ radiative flux changes ΔF versus height z as induced by a disturbance from added ghost heating, SW~~ cirrus~~, LW~~ cirrus, "normal" cirrus with SW,~~ and SW+LW ~~contributions~~contrail cirrus, and 10 % increased CO_2, in the panels from left to right, respectively. Black full lines~~:~~ (i): instantaneous flux; red dashed line~~:~~ (s): adjusted to constant surface temperature; blue dash-dotted line~~:~~ (e): equilibrium over adiabatic surface. The fluxes F are positive if upwards.


p/hPa
0
200
400
600
800
1000
radiative
convective
diffusive
0
1
2
3
ΔT/K
ΔT/K
ΔT/K

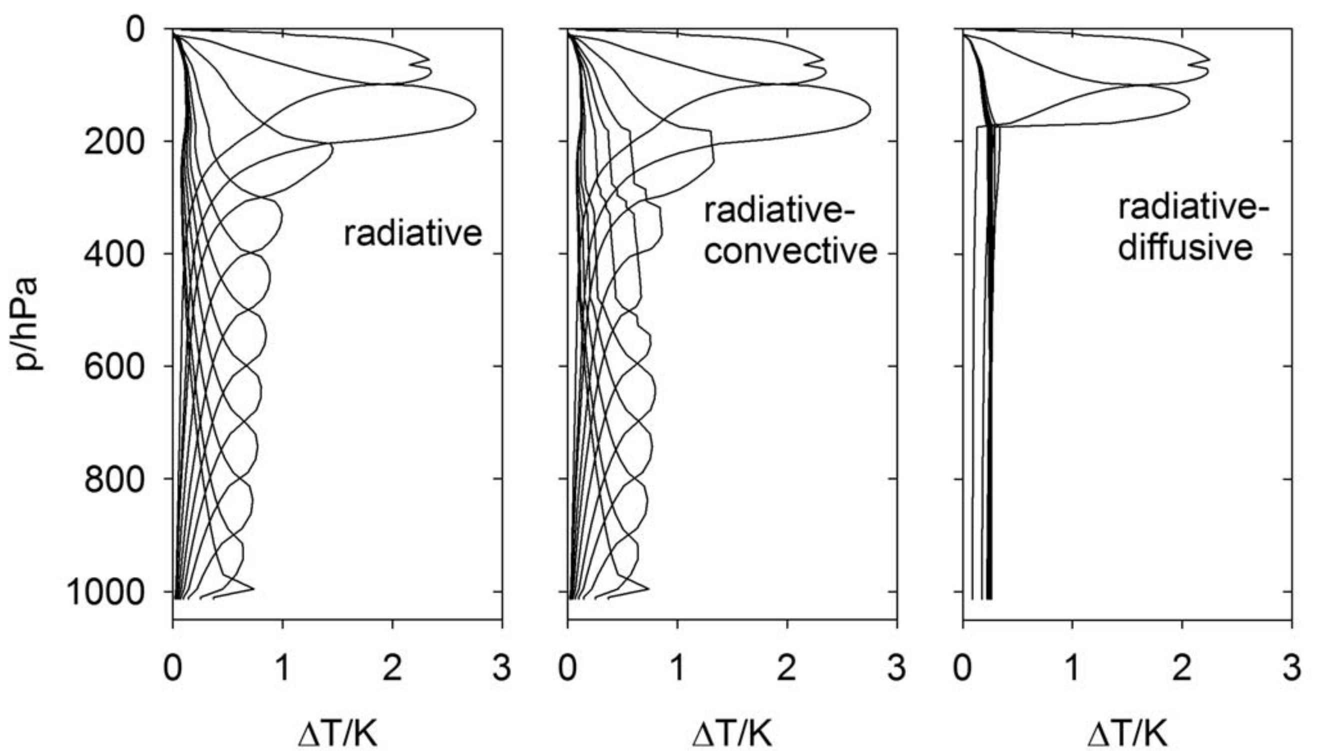
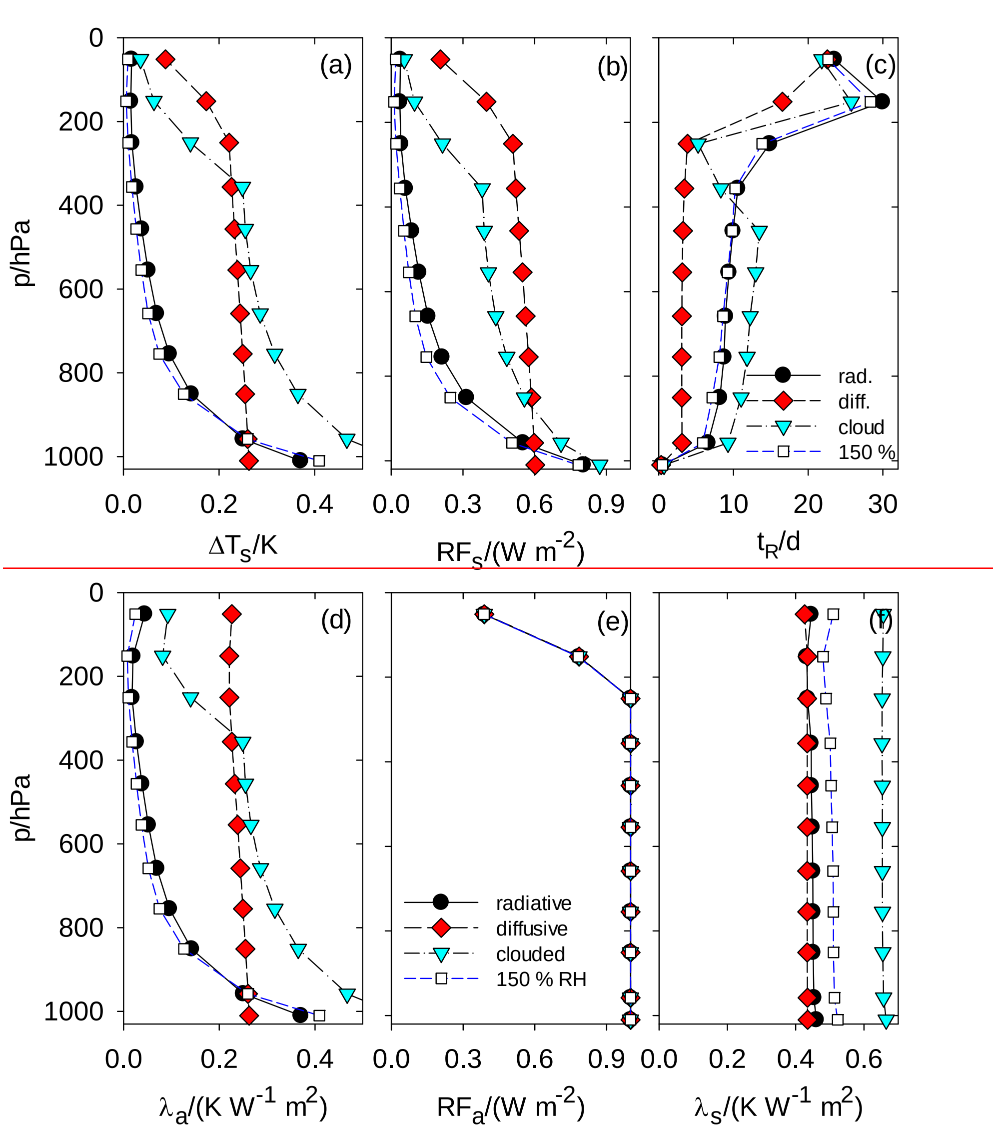


Figure 6: Temperature response~~ profiles~~ versus pressure altitude for layer heating (ghost forcing) with 1 W m^{-2} in ten subsequent 100-hPa pressure layers and ~~at the surface for~~above an adiabatic surface~~ with rapid surface mixing.~~. Left: the “radiative case” with zero turbulent fluxes; middle: the “radiative-convective ~~mixing~~case” with convective adjustment in addition to radiative energy transport; right: ~~for a moderately strong~~ the “radiative-diffusive case” with diffusive mixing ~~κ~~(κ = 100 m^2 s^{-1} ~~constant throughout~~in the troposphere, 0 in the stratosphere) in addition to radiative energy transport.



(a)
(b)
(c)
(d)
(e)
(f)
p/hPa
ΔT_S/K
RF_S/(W m^-2)
t_R/d
rad.
diff.
cloud
150 %
λ_a/(K W^-1 m^2)
RF_a/(W m^-2)
λ_S/(K W^-1 m^2)
radiative
diffusive
clouded
150 % RH

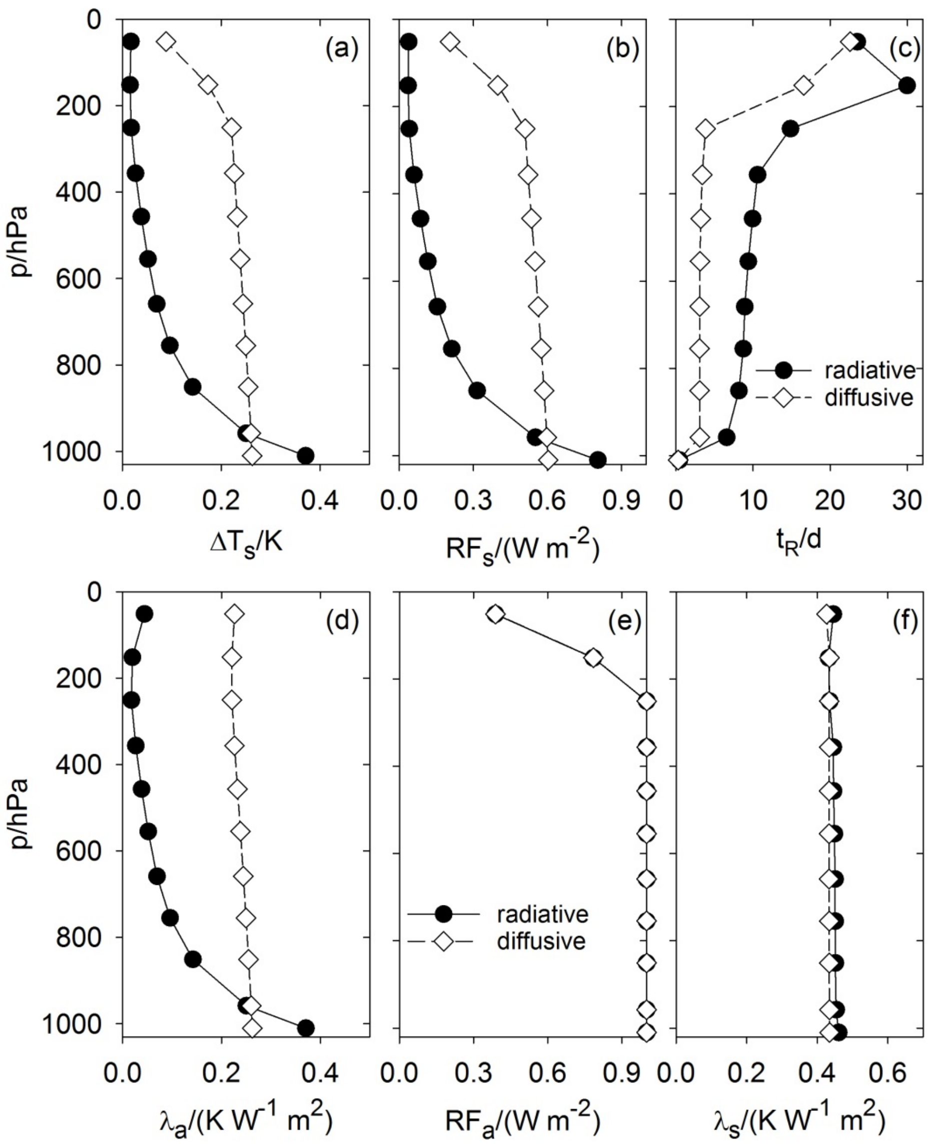
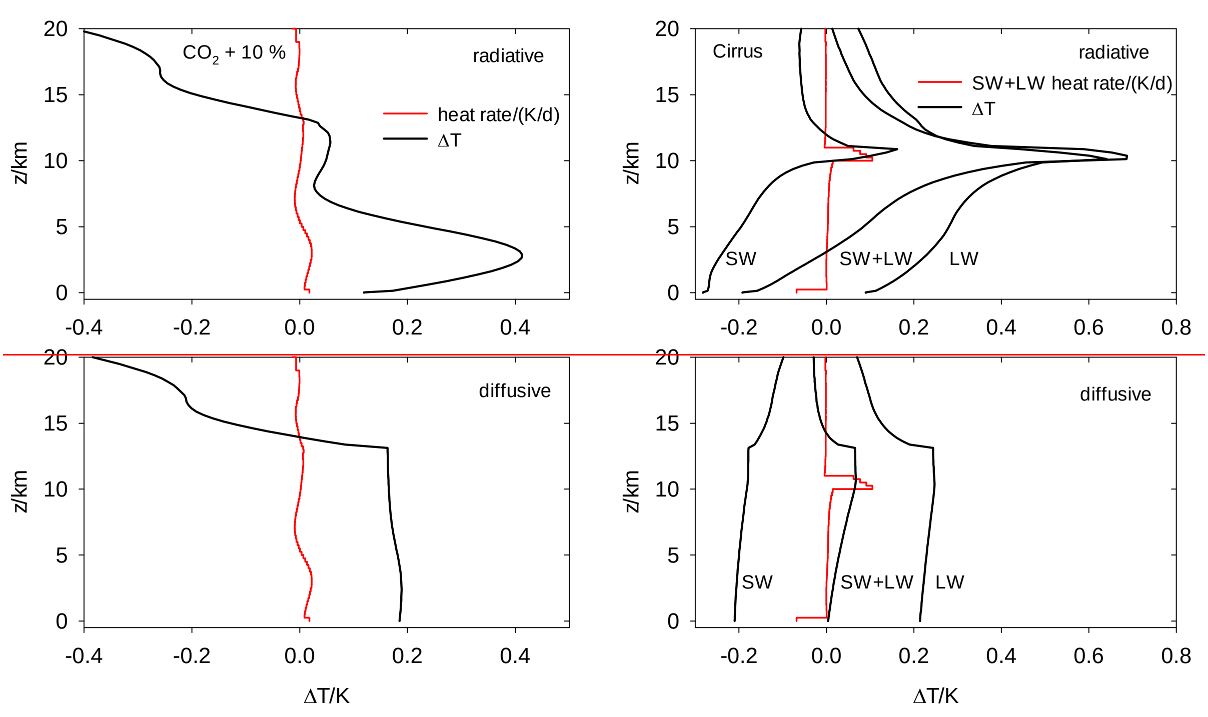


Figure 7: (a) Temperature change at the surface for layer heating versus the layer pressure height~~ in an atmosphere.~~. The ghost forcing corresponds to an RF_i of 1 W m^{-2} at TOA. Black symbols with full ~~lines: model results for radiative equilibrium without mixing; red diamond: with strong diffusive mixing for κ=100 m^2 s^{-1} in the whole troposphere; cyan triangles: with strong diffusive mixing and a 100 % coverage cirrus layer with τ = 3 between 10 and 11 km height; open square with dashed blue ~~line: radiative ~~equilibrium without mixing with 1.5 times enhanced H_2O mixing ratio at all levels in the reference atmosphere~~case; white symbols with dashed line: radiative-diffusive case. (b) Corresponding RF_s values for

fixed T_s. (c) Relaxation time scales $t_R = \Delta T_{layer}/H$. (d) Climate sensitivity parameter $\lambda_a = \Delta T_s/RF_a$ based on stratosphere-adjusted RF_a; (e) RF_a; (f) climate sensitivity parameter $\lambda_s = \Delta T_s/RF_s$ based on effective RF_s.



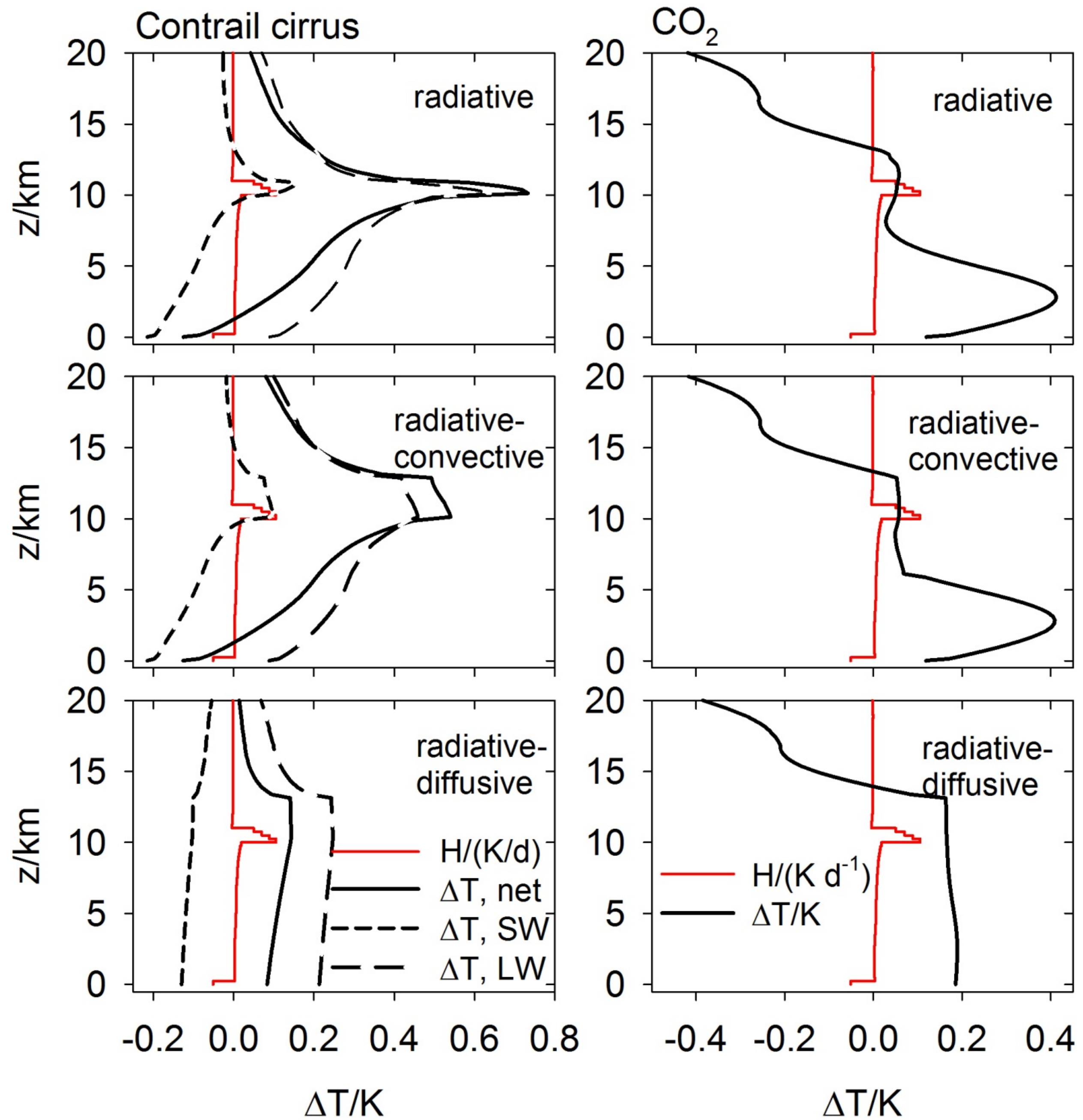
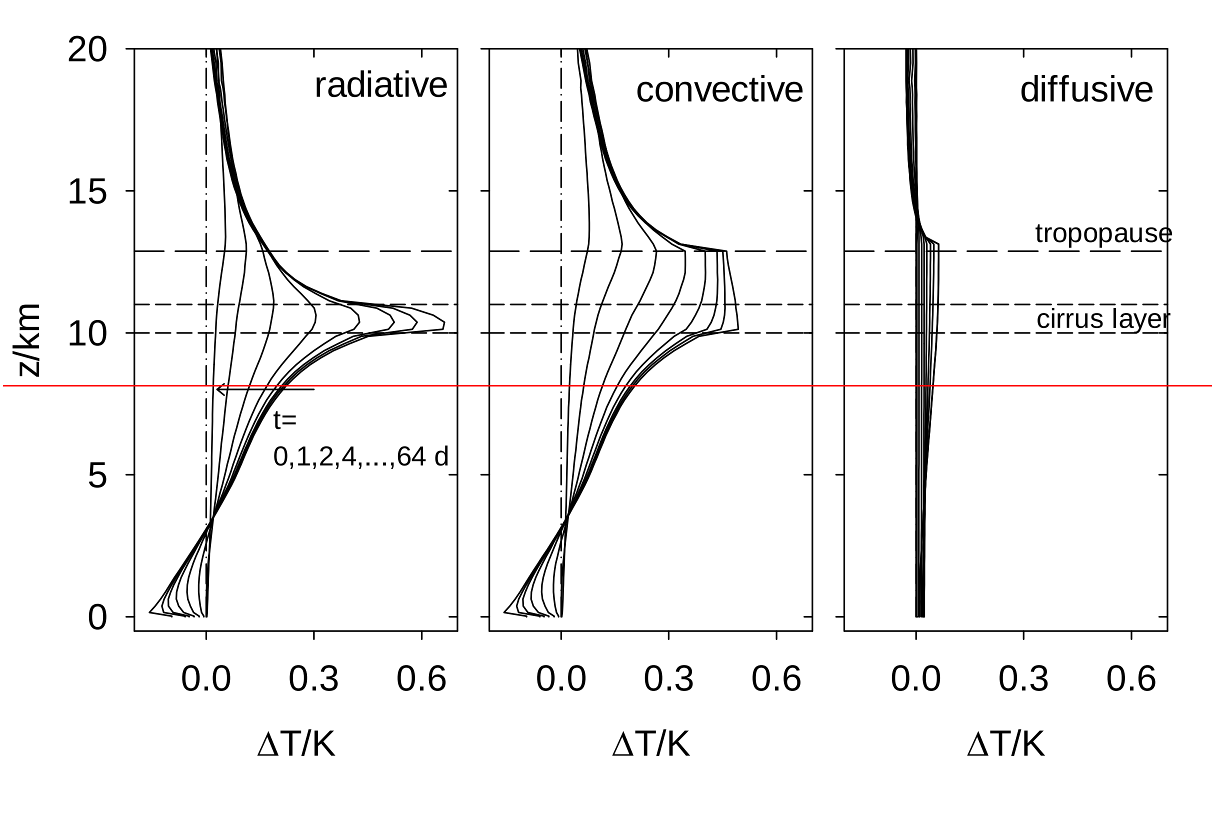


Figure 8~~: Equilibrium temperature~~: Temperature change ΔT in K versus altitude z in km (black lines) for disturbances by CO_2 (left) and by SW, LW and ~~normal~~combined LW+SW contrail cirrus (right~~)~~), in ~~an atmosphere above an adiabatic surface with rapid local mixing at the surface (black line),~~steady state for the radiative~~ equilibrium with zero mixing (top),~~ radiative-convective and ~~with uniform~~ radiative-diffusive ~~tropospheric mixing (bottom).~~cases. The red curves ~~are~~show the net (LW+SW) initial instantaneous heating ratings in K d^{-1}.



radiative
convective
diffusive
z/km
20
15
10
5
0
0.0
0.3
0.6
ΔT/K
ΔT/K
ΔT/K
t=
0,1,2,4,...,64 d
tropopause
cirrus layer

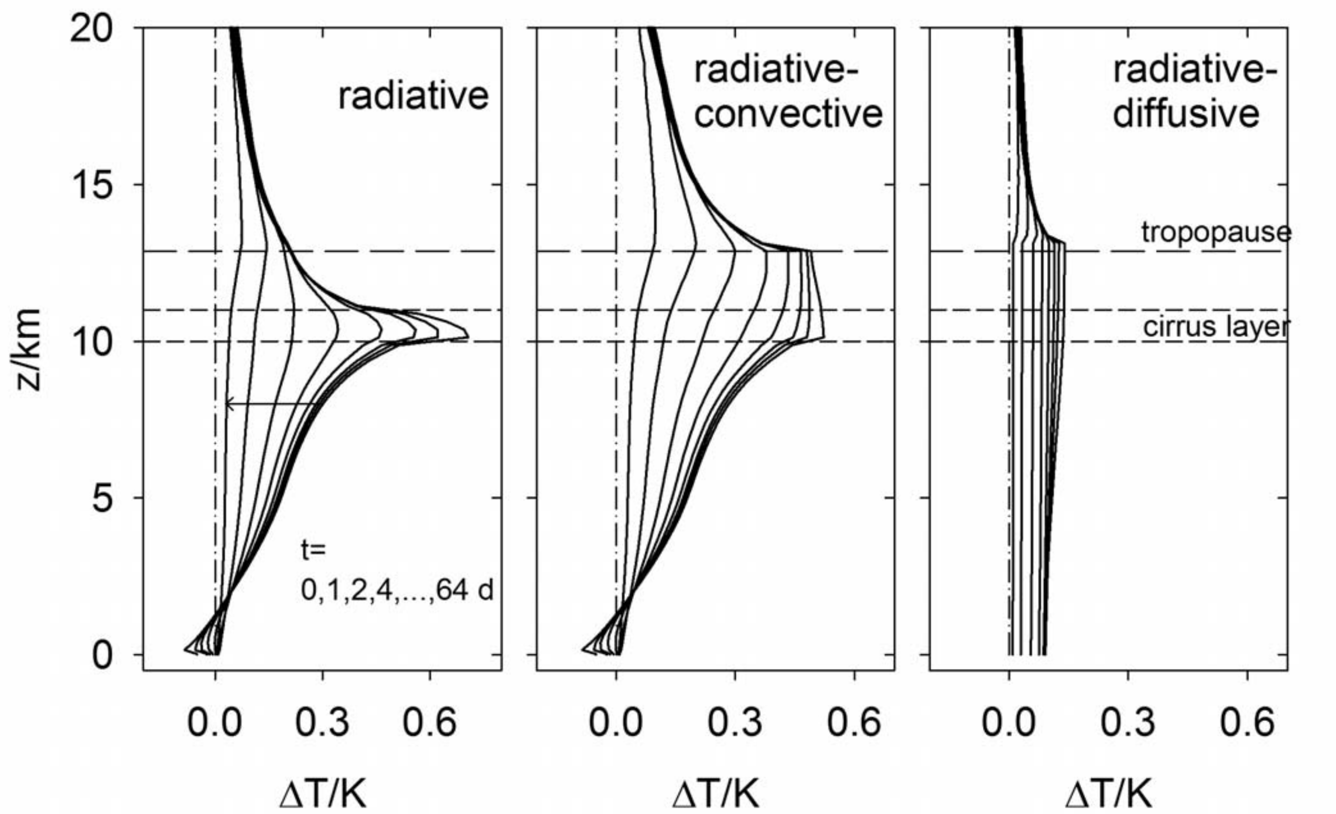
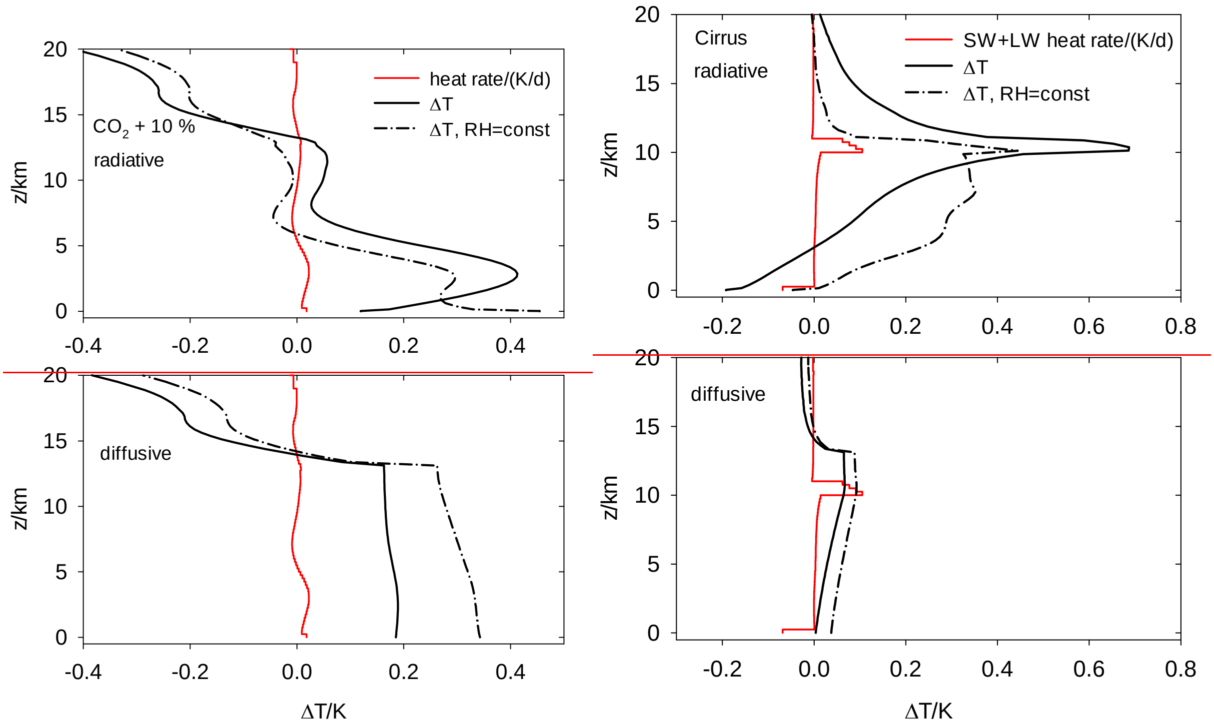


Figure 9. Decay of an ~~initial~~initially steady-state contrail-cirrus-induced temperature increase, at times 0, 1, 2, 4, ..., 64 d after the contrail-cirrus ceased, in three panels for the radiative, radiative-convective and radiative-diffusive ~~mixing~~ cases. Tropopause and ~~cirrus~~contrail layer heights are indicated by dashed lines. ~~The times needed to reach half the initial values are 0.25 d, 22.5 d and 7 d for the temperature at the surface, on average over the troposphere, and in the cirrus layer, respectively, for the radiative case, and shorter for the other mixing cases.~~



~~**Figure 13.** As Figure 11, without (full line) and with (dash-dotted) humidity adapted to constant relative humidity RH (left for CO_2, right for normal cirrus).~~