Investigation of the feasibility of constraints on fast cloud-climate feedbacks from a regional cloud-resolving simulation

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Abstract

Fast cloud-climate feedbacks are an important component of the total cloud-climate feedback, which is still the main source of inter-model spread in climate sensitivity. This study examines fast cloud feedbacks over Central Europe, which are compared to global fast cloud feedbacks by analysing ten global coupled models participating to the Fifth Coupled Model Intercomparison Project (CMIP5). For this purpose, the fast cloud-induced radiative response by the combined Kernel-Gregory method as well as the atmospheric processes by investigating changes in vertical profiles of thermodynamic quantities, which adapt quickly to perturbed CO\textsubscript{2} concentrations, are analysed. Based on the cloud-induced radiative response and even more clearly by the vertical profiles, it shows that Central Europe can be considered as representative for fast cloud feedbacks over global continents since thermodynamic profiles react in similar ways to the perturbation. Despite some robust features across the models for Central Europe such as a decreasing cloud cover in the lower troposphere, the reaction simulated by different models differs for the middle and upper troposphere as well as for the radiative response. Furthermore, by additional sensitivity simulations it is found that the atmospheric processes of fast cloud feedbacks over Central Europe are largely a local atmospheric phenomenon and are rather independent of dynamics effects. These results imply that a high-resolved reference simulation over a limited area might be instructive for an assessment of global cloud-climate feedbacks.
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1 Introduction

1.1 Motivation and outline

As recent studies have shown, an important component of the cloud radiative response to a perturbation of the atmospheric carbon dioxide concentration are fast cloud feedbacks, the reaction of clouds to altered heating profiles (Gregory and Webb, 2008; Block and Mauritsen, 2013). This component accounts for approximately 0.4 to 0.5 W/m$^2$ per doubling of CO$_2$, which contributes about 20% to the modelled total cloud feedback in a model with average sensitivity (Vial et al., 2013; Zelinka et al., 2013). In contrast to the slow cloud feedback, the response of clouds to global changes in surface temperatures, which develops on timescales of years to decades after introducing a forcing, fast cloud feedbacks act within short timescales of the order of days. Therefore, short simulations of climate models are often sufficient to investigate fast feedbacks and their contribution to climate change (Bony et al., 2013).

For simulations over time periods up to one season and over limited domains up to scales of the size of Central Europe, high-resolved simulations by cloud-resolving weather forecast or large eddy models are feasible. It has been demonstrated in many studies that such cloud-resolving simulations can improve our understanding of climate-relevant processes (Zhang et al., 2013; Matsueda and Palmer, 2011). Since such simulations represent cloud processes in a much more reliable way than general circulation models in which clouds are highly parameterised, it is an interesting option to perform sensitivity studies with such large-domain high-resolved models to investigate cloud feedbacks. The large domain and realistic set-up would allow for most of the relevant fast feedbacks to be represented, while the high resolution renders a relatively reliable simulation of cloud feedbacks.

In this masterthesis, data from the 5th Coupled Model Intercomparison Project’s (CMIP5) multi-model ensemble is analysed (Taylor et al., 2011) to investigate fast cloud feedbacks over Central Europe and relate these to the global and continental cloud feedbacks, respectively, to analyse the representativeness of a single region for the global feedbacks. At first, an introduction about the concept of forcing and feedbacks, the important climate feedbacks and fast feedbacks are given in the sections 1.2 to 1.4. After introducing the here used models, experiments and methods
(chapter 2), the radiative response by fast cloud feedbacks (section 3.1) and the underlying atmospheric processes (section 3.2) are analysed. Furthermore, the impact on fast cloud feedbacks over Central Europe by a local CO$_2$ increase versus a global CO$_2$ increase is investigated (section 3.3). In the last section short conclusions are drawn (chapter 4).

### 1.2 Forcing, feedbacks and climate sensitivity

In this section, the basic concept of forcing, feedbacks and climate sensitivity is introduced. It can be well explained by the following radiation equation which is valid for the temporal and global mean:

$$\Delta R(t) = F + \lambda \Delta T_s(t)$$

(1.1)

As a consequence of an external forcing ($F$), this equation describes a linear relation between the time-dependent change of the global mean net radiation fluxes ($\Delta R(t)$) at the top of the atmosphere (TOA) and the time-dependent change of the global mean surface temperature ($\Delta T_s(t)$) multiplied with the so-called climate feedback parameter ($\lambda$). At an unperturbed climate-state ($F=0$), the amount of incoming solar radiation is equal to the amount of reflected solar and outgoing terrestrial radiation, which implies a radiative balance ($\Delta R=0$) and, therefore, surface temperature stays constant ($\Delta T_s = 0$). By introducing a forcing agent ($F>0$), e.g. a CO$_2$ increase, more terrestrial radiation is absorbed by CO$_2$ which leads to a radiative imbalance with $\Delta R>0$ (here the net radiation $R$ is defined as the difference of incoming and outgoing radiation: $R=R^\downarrow - R^\uparrow$). The climate system reacts to this perturbation with an increase of global mean surface temperature ($\Delta T_s > 0$) which implies an enhanced thermal radiation until the atmosphere reaches a new radiative equilibrium. With the temperature increase, different climate constituents as water vapour, surface albedo etc. are altered, which in turn influences the Earth’s radiation budget and therefore the surface temperature. These feedback processes are called climate-feedbacks and can amplify ($\lambda>0$) or dampen the initial forcing ($\lambda<0$). Figure 1.1 shows a schematic overview of forcing and climate feedbacks. The feedback parameter describes the change in radiation per unit change in surface temperature and therefore serves as a measure of the strength of climate feedbacks. The total climate feedback parameter can be split into the individual climate constituents ($x$) under the assumption of linearity and is defined as (Klocke et al., 2011):

$$\lambda = \frac{\partial R}{\partial T_s} = \sum_x \frac{\partial R}{\partial x} \frac{\partial x}{\partial T_s} + \phi(\partial^2) \approx \sum_x \frac{\partial R}{\partial x} \frac{\partial x}{\partial T_s} = \sum_x \lambda_x$$

(1.2)
Forcing, feedbacks and climate sensitivity

Forcing

\( \text{Radiation budget} \)

Surface temperature

Climate constituents

\textit{climate feedbacks}

**Figure 1.1:** Schematic overview of forcing and climate feedbacks

The global increase in surface temperature due to a CO\(_2\) doubling after reaching a new equilibrium is termed as climate sensitivity. Besides the strength of the initial forcing, climate sensitivity depends considerably on climate feedbacks.

1.3 Important climate feedbacks

There are several climate feedbacks acting in the climate system: the most important are namely the Planck-, water vapour-, albedo-, lapserate- and the cloud feedback. All feedbacks can be estimated by an external CO\(_2\) increase within GCMs. Figure 1.2 shows an overview of individual feedback parameters obtained by GCMs analysis from the current IPCC assessment report (Boucher et al., 2013, AR5, Chapter 7), from which the following numbers are taken.

The strongest climate feedback is the so-called Planck feedback, which describes the enhanced radiative effect of thermic radiation due higher surface temperatures. The multi-model mean is simulated with \(-3.2 \pm 0.1 \text{ W/m}^2/\text{K}\) strongly negative and, therefore, counteracts against the positive CO\(_2\) forcing.

The water vapour feedback describes a strong positive feedback process: As a result of warming, the atmosphere can take up more water vapour according to the Clausius-Clapeyron relation. This amplifies warming since water vapour is a strong greenhouse gas. The water vapour feedback is estimated by current GCMs to be \(+1.6 \pm 0.3 \text{ W/m}^2/\text{K}\).

Since the atmosphere responds with a vertical non-uniform warming after increasing CO\(_2\) concentrations, the vertical temperature gradient alters, what is known as lapserate feedback. Thereby, the upper levels of the atmosphere warm up more intensively than the lower levels due to enhanced convection, particularly, in tropics and subtropics (Colman, 2001). Therefore, the upper levels emit more thermal radiation into direction of the outer space compared to the lower atmospheric levels, which imply a cooling effect. In high-latitudes, the lapserate feedback is slightly positive due to weaker convection. However, in the global mean the tropic component predominates (Colman, 2001). The multi-model mean is estimated to be \(-0.6 \pm 0.4 \text{ W/m}^2/\text{K}\).

Often both water vapour and lapserate feedback are added up (similar physical pro-
1 Introduction

cesses) which considerably reduces the model uncertainty range. Then, the combined feedback parameter is assessed to be $+1.1$ ($+0.9$ to $+1.3$) W/m$^2$/K.

Another important climate feedback is the albedo feedback, which plays an extraordinarily role in high-latitudes. Owing to higher surface temperatures, more ice and snow melt. This reduces the surface albedo and increases the surface solar absorption. The multi-model mean is estimated to be $+0.3 \pm 0.1$ W/m$^2$/K.

The cloud feedback describes the cloud response due to higher surface temperatures and consists of several different cloud feedback processes. The radiative effect of these cloud feedback processes strongly depends on the cloud type and cloud properties. The total cloud feedback is assessed to be $0.3 \pm 0.7$ W/m$^2$/K and continues to constitute the main source of inter-model spread in climate sensitivity (Boucher et al., 2013). However, in consideration of additional uncertainties associated with processes that may not have been accounted for in those models, AR5 estimates a broader estimation for the cloud feedback of $+0.6$ (-0.2 to $+2.0$) W/m$^2$/K. Although no clear observational constraint has been found for the total cloud feedback yet, some advances in indentifying robust features for aspects of the cloud feedback

![Figure 1.2: Overview of climate feedback parameters for Planck (P), water vapour (WV), clouds (C), surface albedo (A), lapse rate (LR) and WV+LR for different GCMs. ALL is all except the Planck feedback (Flato et al., 2013).](image)

have been made in recent years. At least, these robust features are consistent across most models and can be physically explained. It has been shown that anvils from deep convection tend to occur at a given temperature, yielding a positive feedback for this cloud type (Hartmann and Larson, 2002). Furthermore, for low-level clouds, there is some indication that the feedback is positive, too, derived from the temporal variability of low-level clouds in the Northern Pacific ocean (Clement et al., 2009) and from the evaluation of a perturbed-physics ensemble simulation with a climate model with observations of the solar cloud radiative effect in subtropical subsidence regions (Klocke et al., 2011). Further robust mechanism are an expansion of the Hadley cell with decreasing cloud cover (Lu et al., 2007) and a poleward shift of the storm tracks (Yin, 2005), both imply a negative cloud feedback. Most of these cloud feedback processes are driven by the general circulation and developed by changes in the surface temperature field (see temperature dependency of $\lambda$; Eq. 1.2) and, therefore, act on timescales of the order of years or decades after introducing a forcing. Besides these slow cloud feedback processes, there are also fast cloud responses owing to higher CO$_2$ concentrations, so-called fast cloud feedbacks 1.

### 1.4 Fast feedbacks

Fast feedbacks are the rapid response of climate constituents which develop without significant changes in the surface temperature field. Compared to the slow feedback processes, fast feedback processes act on short time scales of the order of days or weeks after introducing a forcing. An already well-known rapid response represents the stratospheric cooling: Higher CO$_2$ concentrations immediately enhance absorption of terrestrial radiation in the lower troposphere which leads to a energy loss and thus cooling in the stratosphere (Manabe and Wetherald, 1975). Besides stratospheric cooling, there are also tropospheric rapid responses of climate constituents. By decomposing the climate sensitivity estimated from CMIP5 models into contributions of slow and fast feedbacks and further into their components of the individual climate constituents, Vial et al. (2013) showed that clouds constitute the main contributor of fast feedbacks with about 50%, followed by water vapour and smaller contributions of surface albedo and fast changes in surface temperature. However, models show large differences in how they simulate fast cloud feedbacks, while the inter-model spread is relatively small for the contributions of water vapour and surface albedo (Vial et al., 2013). The main mechanism of fast cloud feedbacks is a CO$_2$ induced radiation effect: In the lower troposphere the temperature is increased due to higher CO$_2$ concentrations which reduces relative humidity and leads

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1 The here termed fast feedbacks are called fast adjustments in the current IPCC AR5.
to sub saturation and therefore partial reduction in cloudiness, which imply a positive feedback (see more in section results).

In addition to this CO$_2$ radiative effect on clouds, a physiological fast response of plants to increase CO$_2$ concentrations influences clouds, too: the plants’ stomata, which considerably control the evaporation rate of the plants’ leaves. Under higher CO$_2$ concentrations, the plants’ stomata are decreased which leads to suppressed evaporation rates of the vegetation over continents and so influence the hydrological cycle and clouds, particularly, in the lower troposphere (Dong et al., 2009). There is observational and modelling evidence of this so-called “CO$_2$ physiological forcing”, however, it cannot be captured well by the radiative forcing concept (Boucher et al., 2009).
2 Methods, models and experiments

In this chapter, the combined Kernel-Gregory method is explained (sections 2.1–2.3), which is an approved method to calculate the cloud radiative effect of fast cloud feedbacks and therefore used in this masterthesis. Furthermore, the here used CMIP5 models and experiments (section 2.4) are introduced as well as specific experiments to investigate the underlying atmospheric processes of fast cloud feedbacks (section 2.5) and the impact of a local CO\textsubscript{2} increase versus a global CO\textsubscript{2} increase on these feedbacks (section 2.6).

2.1 Gregory method

For the specific case of a transient simulation after instantaneous introduction of a forcing agent (such as the abrupt4xCO\textsubscript{2} simulation analysed here), it has been demonstrated that a linear regression of net radiation over temperature yields sensible estimations of climate feedbacks (Gregory et al., 2004; Andrews et al., 2012; Vial et al., 2013). This so-called Gregory method is described by Eq. 1.1, whereby a regression line can be calculated where the slope corresponds to global net feedback parameter. The y-intercept, i.e. the radiation imbalance of this regression line at a time where no surface temperature change is realised yet, corresponds to the radiative forcing at TOA initialised by the external CO\textsubscript{2} forcing. With the splitting of the radiation budget into clear- (R\textsubscript{clr}) and cloudy-sky (R\textsubscript{cld}) components, it was further noted that aspects of the cloud feedback can be analysed by using this regression method (Gregory and Webb, 2008)

2.2 Cloud radiative effect

The cloud radiative effect (CRE) is a simple and fast way of determining R\textsubscript{cld} of the radiation budget at TOA. It is computed by the difference between R in clear-sky (clr) and all-sky (all) conditions which additionally are distinguished in a solar (sw) and terrestrial (lw) radiation component:

\[
CRE = R^{cld} = R^{all} - R^{clr} = R^{all}_{sw} - R^{clr}_{sw} + R^{all}_{lw} - R^{clr}_{lw}
\] (2.1)
Applying the Gregory method to the CRE, it allows to distinguish between the fast and slow component of cloud feedback. As already mentioned before, the slope of the regression line corresponds to the slow feedback parameter. The y-intercept, on the other hand, corresponds to the fast cloud feedback, since the observed cloud radiative effect occurs at a time where no surface temperature has changed yet and therefore agrees to the fast feedback definition (fast and without significant surface temperature changes). However, by using the CRE for the computation of the cloud feedback, a major problem are the so-called cloud masking effects. Not every temporal change of the CRE is caused by the actual change of cloud properties. They can rather be partially produced by cloud masking effects (Zhang et al., 1994; Colman, 2003; Soden et al., 2004). Such a cloud masking effect occurs e.g. for the solar spectrum if the surface albedo is altered and the clouds remain unchanged. For a reduction in surface albedo, the clear-sky term \( R_{sw}^{clr} \) increases, which causes a decrease in CRE without any change in cloud properties. Since the surface albedo is a function of \( T \), a dependency between CRE and \( T \) would be noticed. But such a dependency caused by cloud masking effects is not recognised as cloud feedback.

### 2.3 Combined Kernel-Gregory method

In order to quantify the effects by cloud masking and to correct the CRE accordingly, the radiation effects induced by the individual climate constituents in all- and clear-sky conditions have to be calculated. For this purpose, first \( \Delta R \) is decomposed at the TOA into the contributions by the individual climate constituents of water vapour (W), surface albedo (A), clouds (C), \( \text{CO}_2 \) and temperature (T):

\[
\Delta R \approx \Delta R_W + \Delta R_A + \Delta R_C + \Delta R_{\text{CO}_2} + \Delta R_T
\]  

(2.2)

In order to quantify these radiation effects, there are different approaches (e.g. Klocke et al., 2013). If two equilibrium simulations are available, often offline radiative transfer simulations are performed in the “partial radiative perturbation” method (Colman, 2003; Tomassini et al., 2013). A simplified version of this are the “radiative kernels” (Soden et al., 2008) where a unit perturbation rather than the actual one between two climate states are applied. The radiative kernel \( K_x \) can be formulated for each climate constituents (x) as:

\[
K_x = \frac{\partial R}{\partial x}
\]  

(2.3)
Then, the radiative effect at the TOA induced by the climate constituents is obtained by multiplying the kernel with the respective monthly mean change in $x$ between the control and perturbed simulation $\Delta x$:

$$\Delta R_x = K_x \Delta x$$  \hspace{1cm} (2.4)

Unfortunately, there is no cloud kernel available, since a unit perturbation of clouds is difficult to realize due to their complexity. Instead of calculating the cloud feedback directly, the kernels can be used to quantify cloud masking effects to correct the CRE accordingly. For this, in analogy to the CRE kernels in all- and clear-sky conditions are calculated. The differences between the respective all-sky ($K^\text{all}_x$) and clear-sky ($K^\text{clr}_x$) kernels multiplied with $\Delta x$ account for the radiative effects by cloud masking. Corrections are necessary for surface albedo, temperature, water vapour and CO$_2$ following Block and Mauritsen (2013) the single correction terms are:

$$\Delta R^\text{mask}_A = (K^\text{all}_A - K^\text{clr}_A) \Delta A$$  \hspace{1cm} (2.5)

$$\Delta R^\text{mask}_T = \Delta (K^\text{all}_{T_s} - K^\text{clr}_{T_s}) \Delta T_s - \int_{p_s}^{0} [(K^\text{all}_T - K^\text{clr}_T) \Delta T] dp$$

$$\Delta R^\text{mask}_W = \int_{p_s}^{0} [(K^\text{all}_W - K^\text{clr}_W) \Delta \ln(q)] dp$$

$$\Delta R^\text{mask}_{\text{CO}_2} = (K^\text{all}_{\text{CO}_2} - K^\text{clr}_{\text{CO}_2}) \Delta \log_2(\text{CO}_2)$$

Thereby, the albedo kernels are three dimensional (time, latitudes and longitude) and therefore have to be only applied to the albedo change in the surface layer. Water vapour and temperature kernels have an additional height dimension and thus have to be applied for each atmospheric layer and subsequently vertically integrated from the surface ($p=p_s$) to the TOA ($p=0$). Additionally, for the surface temperature there are separate kernels. The CO$_2$ kernels are only applied once during the forcing. To obtain the kernel-corrected CRE ($\Delta R_C$) then, the correction terms are subtracted from the CRE:

$$\Delta R_C \approx \Delta CRE - \Delta R^\text{mask}_A - \Delta R^\text{mask}_T - \Delta R^\text{mask}_W - \Delta R^\text{mask}_{\text{CO}_2}$$  \hspace{1cm} (2.6)

Now, the CRE can be replaced by the $\Delta R_C$ in the Gregory method, what is known as combined Kernel-Gregory method, to obtain more reliable results for fast cloud feedbacks.
In this masterthesis, the combined Kernel-Gregory method is applied to the annual-mean changes of $\Delta R_C$ and $\Delta T_s$ to yield the radiation effect of fast cloud feedbacks by the y-intercept.

## 2.4 Models and experiments

The analysis of fast cloud feedbacks is applied to ten coupled global climate models as submitted to the CMIP5 archive. Monthly mean values of the CMIP5 model outputs are used (for a complete list see Table 2.1). In order to obtain a sufficiently strong fast cloud feedback signal, the idealised abrupt4xCO$_2$ experiment (instantaneous quadrupling of the atmospheric carbon dioxide concentration over pre-industrial levels; Taylor et al., 2011) in comparison to the pre-industrial control run (piControl) at unperturbed CO$_2$ concentrations are used. For the calculation of the radiative response by fast cloud feedbacks the combined Kernel-Gregory technique is applied to the first 25 years of a single realisation of the abrupt4xCO$_2$ minus the piControl experiment. The radiative kernels of Shell et al. (2008) are applied to all ten models analysed.

### Table 2.1: CMIP5 models and purpose of analysis.

<table>
<thead>
<tr>
<th>Modelling Center</th>
<th>Modell version</th>
<th>Radiative response</th>
<th>Vertical profiles</th>
</tr>
</thead>
<tbody>
<tr>
<td>BCC, China</td>
<td>bcc-csm1-1</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>CCCma, Canada</td>
<td>CanESM2</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>CNRM-CERFACS, France</td>
<td>CNRM-CM5</td>
<td>x</td>
<td></td>
</tr>
<tr>
<td>IPSL, France</td>
<td>IPSL-CM5A-LR</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>MIROC, Japan</td>
<td>MIROC5</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>MOHC, UK</td>
<td>HadGEM2-ES</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>MPI-M, Germany</td>
<td>MPI-ESM-LR$^2$</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>MRI, Japan</td>
<td>MRI-CGCM3</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>NCAR, USA</td>
<td>CCSM4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NOAA-GFDL, USA</td>
<td>GFDL-CM3</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

## 2.5 12-member ensemble

When investigating the atmospheric processes beyond the radiative response of fast cloud feedbacks (see section 3.2), vertical profiles of different meteorological quantities as potential temperature, specific humidity and cloud cover are calculated.

$^2$For MPI-M the radiative responses are calculated by CMIP5 data, whereas the data of the vertical profiles are obtained by simulations with MPI-ECHAM6 with fixed-SST and a CO$_2$ concentration change within the 4xCO$_2$ forcing experiment from 367 to 1470 ppm, while the CO$_2$ concentration within the CMIP5 models is increased from 285 to 1140 ppm.
Since fast cloud feedbacks act within short timescales of days, it is sufficient to analyse the first month of the simulation after the perturbation in comparison to the corresponding unperturbed time period. To obtain a good statistical representativeness, and to remove seasonality, a 12-member ensemble is used in which the first member is initialised in January, the second in February and so on (Taylor et al., 2011). In the end, the mean over all months is calculated. In a similar approach, Kamae and Watanabe (2013) analysed results from their MIROC5 model. Unfortunately, not every model within CMIP5 provides all multiple realisations which are necessary for the 12-member ensemble. Hence, the vertical profiles are only calculated from a subset of seven models (see Table 2.1). The global climate model of the Max-Planck-Institute (MPI) for Meteorology (MPI-ESM-LR) did not submit a 12-member ensemble initialised at different times of the year to CMIP5. Since this model is used here, however, for a specific analysis, additional simulations (a one-year control experiment after three months of spin-up at present-day CO$_2$ conditions, and twelve one-month experiments at quadrupled CO$_2$ concentrations starting each from a different month of the control experiment) with the newer version of its atmospheric component (ECHAM6 Stevens et al., 2013) at a resolution of T63L47 with prescribed sea surface temperatures are performed. Since the focus here is on the fast cloud feedbacks, these AMIP-type experiments with prescribed sea surface temperatures and sea ice contents can be considered as representative to what a fully coupled model would yield. The ocean is expected to react to the CO$_2$ perturbations at time scales much longer than the ones analysed here.

2.6 MPI-ECHAM6 set-up

The impact on the fast cloud feedback over Central Europe by a local CO$_2$ increase versus a global CO$_2$ increase (see section 3.3) is determined by conducting simulations with MPI-ECHAM6. The area Central Europe is defined here as corresponding to the domain of the German Weather Service regional-area forecast model COSMO-DE (4°E–14.5°E, 45°N–56.5°N). Rather than increasing CO$_2$ globally, as in the simulations described above, in this second step, the CO$_2$ concentrations are quadrupled in the Central European region but kept fixed at the reference concentration elsewhere on the globe.

$^3$The control simulation and model version used for the MPI-ESM-LR contribution to CMIP5 unfortunately were not available to me so an investigation using a coupled model would have implied the need for a new spinup of a coupled control simulation.
3 Results

3.1 Radiative response by fast cloud feedbacks

This section provides an overview of the radiative response by fast cloud feedbacks analysed for ten models with the combined Kernel-Gregory method. In the first step, a linear regression analysis by the annual-mean changes of $R_C$ and $T_s$ as global mean values are calculated for all ten models (Figure 3.1). In agreement with previous studies (e.g. Vial et al., 2013), all models exhibit a positive y-intercept which implies a positive global fast cloud feedback. In spite of a consistent sign between the models, they show a large inter-model spread, which ranges from 0.35 to 1.89 W/m$^2$ with a multi-model mean value of 0.96 W/m$^2$. Furthermore, the degree of linear correlation ($R^2$) between $\Delta R_C$ and $\Delta T_s$ differs a lot between the models: MIROC5 and CCSM4 show almost no correlation while the models MRI-CGCM3, BCC-CSM1.1 and CNRM-CM5 show higher correlations and MPI-ESM-LR, IPSL-CM5A-LR, CanESM2 and GFDL-CM3 even simulate a strong and clear response of $\Delta R_C$ to higher values of $\Delta T_s$. The noise is considerable reduced by averaging over all models (see multi-model mean).

In order to obtain a global distribution of fast cloud feedbacks, it is possible to calculate a global distribution of y-intercepts by regressing the annual-mean change of $\Delta R_C$ at every grid point against the global and annual-mean change of $\Delta T_s$ (Gregory and Webb, 2008) (Figure 3.2). It shows that the majority of models simulates a more pronounced positive effect over the land areas where surface temperatures react on faster timescales than over the ocean (Boucher et al., 2013). In particular, a strong positive fast cloud feedback is found in most models over Central Europe, North and South America, Russia and over the warm pool around Indonesia.

These results are confirmed in more detail by regressing $\Delta R_C$, first as an area-average over all land areas (Figure 3.3) and in a second step as an area-average over the specific region Central Europe (Figure 3.4) against the global mean of $\Delta T_s$. Compared to the global (land-and-ocean) mean fast cloud feedback (Figure 3.1), most models exhibit a larger fast cloud feedback over the global land areas (multi-model mean of 1.45 W/m$^2$) with a simultaneous increase in inter-model spread, which ranges from 0.75 to 2.97 W/m$^2$. 
Figure 3.1: Application of the Kernel-Gregory method for all then models: Regression of the global and annual-mean change of the kernel-corrected cloud radiative effect ($\Delta R_C$) against the global and annual-mean change of the surface temperature ($\Delta T_s$). The analysis is applied to the first 25 years of the abrupt4xCO$_2$ forcing simulation minus the piControl experiment.

Furthermore, most models show a similar $R^2$ for the regression analysis over the global land areas compared to the global mean while $R^2$ is different only for few models (e.g. increasing for MIROC5 and decreasing for MPI-ESM-LR).
3 Results

For the domain Central Europe, models show much more variability and noise, i.e. considerably smaller $R^2$, and larger fast cloud feedback values with also larger inter-model spread (2.05 to 9.5 W/m$^2$) with a multi-model mean of 5.42 W/m$^2$ (Figure 3.4).

Figure 3.5 shows the area-average of fast cloud feedbacks over the Central-European domain vs. global land areas for the individual models. It is found that models with a higher cloud-induced radiative response over Central Europe tend to have a higher response over the global land areas. However, the values are about four times smaller when averaging over all land areas compared to averaging just over Central Europe. Still it may be concluded that the radiative response over Central Europe seems to behave qualitatively similar as it does over global land areas in individual models.

Figure 3.2: Geographical distributions of the fast cloud-induced radiative response [W/m$^{-2}$] for all ten models, multi-model-mean and multi-model standard deviation. The responses are obtained by the y-intercepts from the regression analysis of the combined Kernel-Gregory method, which is applied for the abrupt4xCO2 forcing simulations minus the piControl simulations.
3 Results

Based on the Figures 3.2 and 3.5, it can be concluded that at a first glance Central Europe looks qualitatively representative for fast cloud feedbacks over global land areas. Further investigations of the representativeness of Central Europe for fast cloud feedbacks and in particular the underlying atmospheric processes are dealt with in the next section.

\[\Delta R_{\text{c, global land areas}} [\text{Wm}^{-2}]\]

\[\Delta T_{\text{s, global}} [\text{K}]\]

**Figure 3.3:** As Figure 3.1, whereby on the y-axis the mean over all continents is plotted.
Figure 3.4: As Figure 3.1, whereby on the y-axis the area-average of the Central-European domain is plotted.
Figure 3.5: Area-average of radiative response of fast cloud feedbacks (FCF) over Central Europe (x-axis; obtained by the y-intercepts from Figure 3.4) and over global land areas (y-axis; obtained by the y-intercepts from Figure 3.3) for all ten models. The error estimation for the y-intercepts is based on the regression standard error.
3 Results

3.2 Changes of vertical profiles due to fast cloud feedbacks

To understand the atmospheric processes beyond the radiative responses, the multimodel mean changes due to a 4xCO$_2$ forcing relative to the control run for the mean of the 12-member ensemble as vertical profiles of the following meteorological quantities are shown: potential temperature ($\theta$), relative humidity (RH), specific humidity (q) and cloud cover (cl) (Figure 3.6). For the global (land-and-ocean) mean (Figure 3.6, left column) a higher CO$_2$ concentration induces a slight warming of all atmospheric layers with a maximum at about 850 hPa. Consequently, relative humidity decreases and thus widely correlates with the temperature profile. However, higher temperatures at ground levels usually result in an increased evaporation rate and thus enhanced specific humidity, which in turn tends to increase relative humidity. In the sum, the reduction of relative humidity by the temperature increase predominates and leads to sub-saturation and thus a decrease of low and middle clouds, which implies a positive-sign (warming) fast cloud feedback. The basic principle of these atmospheric processes are similar between land (Figure 3.6, middle column) and ocean areas (not shown) although there are some small differences: The temperature response over ocean areas are weaker, while the increase of specific humidity is stronger. Both effects lead to a lower decrease of relative humidity and cloud cover over oceans compared to land areas.

How do the vertical profiles change for Central Europe? The changes of potential temperature, relative humidity and cloud cover indicate a similar pattern for Central Europe as were found for the global land areas (Figure 3.6, rightmost column). A main difference between Central Europe and the global land areas in the multimodel mean is the specific humidity in the lower levels, which is decreasing for Central Europe and increasing for the global land areas. Nevertheless, the response in RH and subsequently cloudiness is dominated by the warming and as such this discrepancy does not influence the results for fast cloud feedbacks much. However, the vertically more uniform warming over Central Europe leads to a drying in the upper troposphere in this region, which is not found when averaging over global continents. In consequence, the cloud cover over Central Europe is decreasing above 500 hPa while it is slightly increasing for global land areas.

When considering subsets of the 12-member ensemble, each containing three months (i.e., one season), the changes become much noisier. Still, for temperature, RH and cloud cover, they still qualitatively look similar to the annual-average results.
Figure 3.6: Changes of potential temperature, relative humidity, specific humidity and cloud cover for the multi-model mean calculated for the global (land and ocean) mean, global land mean and Central European mean as vertical profiles. The responses are calculated by the 4xCO$_2$ forcing simulation minus the control simulation. The black solid line represents the mean of the 12 member ensemble, while the coloured lines (in the right column only) are subsets of the 12-member ensemble, each for a three-month average, i.e. for one season and for the multi-model ensemble average. The blue-shaded area around the multi-model mean indicates the standard deviation of both the multi-model mean and the model’s 12-member ensemble.
Figures 3.7 to 3.10 show the responses to quadrupling CO$_2$ as simulated by the individual models. The profiles of induced changes in potential temperature in all models are similar between global averages and land-only global averages, but with smaller amplitudes when also considering the ocean (Figure 3.7). In all models but BCC CSM-1 the temperature response over Central Europe is qualitatively and even quantitatively representative for the global-land response. This translates into similar findings for RH (Figure 3.8). In three models (IPSL-CM5A-LR, MRI-CGCM3 and HadGEM2-ES) the response is very similar over Europe and global continents, even quantitatively. BCC-CSM1-1 shows a small and thus noisy response in RH. The other three models (MIROC5, CanESM2, and MPI-ECHAM6) agree between Central Europe and global continents in the lower troposphere (MIROC5 and CanESM2 with stronger drying over Europe), but show a different behaviour in the upper troposphere. Specific humidity again looks smoother and more similar between Central European and global continental averages, except for MIROC5 and BCC-CSM1-1 that show a drying in specific humidity over Europe vs. wettening globally (Figure 3.9). In terms of cloudiness, a common feature across all models is the decreased cloud cover in the lower troposphere with a range from 1–2% over Central Europe (Figure 3.10). This is in accordance with the cloud response over the global land areas, albeit the magnitude of the changes are stronger. Over Central Europe, four models (MIROC5, CanESM2, IPSL-CM5A-LR, and MPI-ECHAM6) simulate a cloud cover decrease for all atmospheric layers. For MIROC5 this is larger than for global land, but qualitatively in agreement, and for IPSL-CM5A-LR the response is similar to the global-land one. CanESM2 and MPI-ECHAM6, however, show increases in cloudiness when averaging over global land. BCC-CSM1-1 shows small, and noisy changes but broadly similar between Europe and global land. The two models (MRI-CGCM3 and HadGEM2-ES) that show decreases in upper-tropospheric cloud cover over Europe also show a same signal over global land, although HadGEM2-ES shows a response different in sign in the middle troposphere.

In summary, there are substantial differences in how models simulate global and regional feedbacks to increased CO$_2$. However, mostly the results over Central Europe seem at least qualitatively representative for what the models simulate globally and subsequently better insights into what occurs in the particular region may be instructive for a global cloud feedback assessment.
Figure 3.7: As Figure 3.6 but for each individual model and potential temperature only. The simulations of MPI-ECHAM6 are performed with fixed-SST and a CO₂ concentration change within the 4xCO₂ forcing experiment from 367 to 1470 ppm, while the CO₂ concentration within the CMIP5 models is increased from 285 to 1140 ppm.
Figure 3.8: As Figure 3.7 but for changes of relative humidity.
Figure 3.9: As Figure 3.7 but for changes of specific humidity.
Figure 3.10: As Figure 3.7 but for changes of cloud cover.
3 Results

3.3 Local CO$_2$ increase versus global CO$_2$ increase

How strongly do dynamics influence fast cloud feedbacks and what are the implications for the representativeness of Central European cloud feedbacks? To analyse this, additional simulations with MPI-ECHAM6 are conducted with CO$_2$ quadrupling over Central Europe only, in which the CO$_2$ concentrations of the remaining areas are kept fixed at control values (see section 2.6). Besides estimating the dynamical component of fast cloud feedbacks over Central Europe this approach also allows to investigate to what extent a reference sensitivity simulation using a cloud-resolving model over a limited area domain, in which only the CO$_2$ concentration within the limited area can be increased but which would still be driven with present-day meteorology at its boundaries, might be instructive for fast cloud feedbacks. Figure 3.11 shows the vertical profiles for Central Europe of the same meteorological quantities as in Figure 3.6, but comparing the local versus global CO$_2$ increase. Interestingly, for most quantities, only small differences are found: For the local CO$_2$ increase the potential temperature increase is somewhat smaller, and negative in the tropopause region. However, the relative humidity decrease is larger due to a decrease of specific humidity, opposite to the increase simulated when quadrupling CO$_2$ globally. Therefore, the simulated reduction in cloud cover is more pronounced. Despite these differences, the results show that the vertical profiles evolve similarly for both, global and regional CO$_2$ quadrupling. Thus, the fast response to the CO$_2$ perturbation largely seems to be a local phenomenon and is relatively independent of dynamic effects, they are rather driven by the thermodynamic component. As a consequence, Central Europe seems to be a suitable region for a comparison between simulations of CMIP5 models and simulations from cloud-resolving models. One general issue when comparing simulations of limited domains between GCMs and high-resolved models arises due to the given boundary conditions in the latter ones which might affects the results near the boundaries and therefore should be considered in the analysis (e.g. Seifert et al., 2012).
Figure 3.11: Changes of vertical profiles as in Figure 3.6 for MPI-ECHAM6 AMIP-type simulations for a global CO₂ increase (left column) and a local CO₂ increase over Central Europe (right column). The local CO₂ increase is implemented by quadrupling the CO₂ over Central Europe, whereby the remaining areas’ CO₂ concentrations are kept fixed at the reference concentration. The blue-shaded area around the model mean indicates the standard deviation of the model’s 12-member ensemble.
Fast cloud feedbacks as simulated in the idealised experiments of the CMIP5 multi-model ensemble have been analysed globally, and for Central Europe. It was found that the fast cloud-induced radiative response over Central Europe behaves qualitatively similarly as over global land areas. Furthermore, the investigation of the underlying atmospheric processes by assessment of changes in vertical profiles of different meteorological quantities, which quickly adapt to a new CO$_2$ concentration confirmed this finding to a large extent: The vertical profiles over Central Europe in general are altered in a similar way as for the global land areas to the same CO$_2$ perturbation. Nevertheless, comparing the individual models, a large inter-model spread exists: Despite some robust features across the models for Central Europe as a decreased cloud cover in the lower troposphere, models differ for the middle and upper troposphere as well as for the radiative response. Finally, specific simulations are performed to assess whether a regional-only increase in CO$_2$ still yields similar results in terms of the perturbation, which largely is found to be true in the ECHAM6 mode. It may be concluded that fast cloud feedbacks seem to be a local, thermodynamically-driven phenomenon. These results imply that a reference simulation with quadrupled CO$_2$ concentrations using a limited-area cloud-resolving model, when performed over sufficiently long times (at least one season), even if driven by present-day boundary conditions (potential impacts near the boundaries should be considered when compare with GCMs), can be instructive to constrain global fast cloud feedbacks and, in consequence, global climate sensitivity.
Declaration

I, Philipp Kühne, matriculation number 2143111, hereby confirm that I have written the present masterthesis with the working title

Investigation of the feasibility of constraints on fast cloud-climate feedbacks from a regional cloud-resolving simulation

independently and without use of others than the indicated aids. I further confirm that the thesis has not been submitted to other previous examinations. I would like to point out that parts of the thesis have been used for a scientific article with the same title which is submitted to the Journal of Advances in Modeling Earth System (JAMES) on August 22, 2014.

Leipzig, September 30, 2014

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Bibliography


