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Cloud radiative forcing of subtropical low level clouds in global models

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Abstract Simulations of subtropical marine low clouds and their radiative properties by nine coupled ocean-atmosphere climate models participating in the fourth assesment report (AR4) of the intergovernmental panel on climate change (IPCC) are analyzed. Satellite observations of cloudiness and radiative fluxes at the top of the atmosphere (TOA) are utilized for comparison. The analysis is confined to the marine subtropics in an attempt to isolate low cloudiness from tropical convective systems. All analyzed models have a negative bias in the low cloud fraction (model mean bias of –15%). On the other hand, the models show an excess of cloud radiative cooling in the region (model mean excess of 13 W m^{-2}). The latter bias is shown to mainly originate from too much shortwave reflection by the models clouds rather than biases in the clear-sky fluxes. These results confirm earlier studies, thus no major progress in simulating the marine subtropical clouds is noted. As a consequence of the combination of these two biases, this study suggests that all investigated models are likely to overestimate the radiative response to changes in low level subtropical cloudiness.

Keywords Low clouds · Cloud radiative forcing · Global climate models · Stratocumulus · Cloud feedback · ERBE · ISCCP · IPCC

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1 Introduction

Clouds are of great concern for the climate of the Earth, due to their interaction with the radiation budget. On one hand, they shield the surface from solar radiation (the albedo effect) and thereby have a cooling effect on the climate. On the other hand, clouds have a warming effect since they prevent terrestrial longwave radiation to escape to space (the greenhouse effect of clouds). Thus, depending on the characteristics of the clouds (height, optical thickness etc.), they have the potential to either warm or cool the surface. On a global average clouds have a cooling effect on climate (Ramanathan et al. 1989).

A convenient way to measure how clouds affect the Earth's radiation budget is the cloud radiative forcing (CRF), defined at the top of the atmosphere (TOA) (Charlock and Ramanathan 1985; Coakley and Baldwin 1984). The CRF is commonly divided into a longwave (CRF_{LW}) and a shortwave component (CRF_{SW})

$$CRF_{LW} = LW\uparrow_{clear-sky} - LW\uparrow_{all-sky} \tag{1}$$

$$CRF_{SW} = SW\uparrow_{clear-sky} - SW\uparrow_{all-sky} \tag{2}$$

$$CRF_{net} = CRF_{LW} + CRF_{SW} \tag{3}$$

where the subscript clear-sky implies the radiant flux when no clouds are present and the subscript all-sky implies the radiant flux when clouds could be present. The CRF_{LW} is in general positive since clouds tends to decrease the effective radiation temperature and CRF_{SW} is in general negative due to the albedo effect. CRF and its components depend strongly on the properties of the clouds (height, optical depth etc.) and there is no clear-cut relation between CRF and the cloud cover (Hartmann et al. 1992), i.e. a given CRF can be produced by many different cloud configurations. In general, high, optically thin clouds have a heating effect on climate (CRF > 0) since the albedo effect (CRF_{SW}) is small but the greenhouse effect (CRF_{LW}) due to the low cloud-top temperature is large. For optically thick medium-high and high clouds the counter-acting effects tend to balance (CRF ~ 0), while low clouds in general are optically thick whilst not altering the outgoing longwave radiation much. Low clouds thus tend to have a net cooling effect on the climate (CRF < 0).

Different clear-sky conditions (i.e. surface albedo, aerosol load and atmospheric absorption) also alter the amount of CRF (Li and Trishchenko 2001). As an example, two identical cloud setups—one over the ocean and one over a snow covered surface would result in significantly different cloud radiative forcing, due to the difference in surface albedo.

Hartmann et al. (1992) showed that variations in low clouds on a global average basis, are the largest contributors to the variations in the Earth's radiation budget. In fact, an absolute increase in low cloudiness by only 5% could cancel out the radiative forcing due to a doubling of CO_2 (Slingo 1990).

The response of cloudiness to radiative forcing associated with variations in greenhouse gases, anthropogenic aerosols or other processes that impact the climate system is of great importance for the climate sensitivity, for an extensive review on the cloud-climate feedback we refer to Stephens (2005). Much of the inter-model variation of future climate projections is associated with how cloudiness in the general circulation models (GCMs) responds to a particular radiative forcing (Cess et al. 1990; Soden and Held 2006; Webb et al. 2006). Recently, Bony and Dufresne (2005) have shown that the largest differences in models' cloud responses to an increase in greenhouse gases are found in regions associated with large-scale subsidence (i.e. the subtropics).

Hitherto, most GCM evaluations and intercomparisons of cloudiness and its influence on the radiation budget have been made on a global- or zonal-mean basis (e.g. Weare et al. 1996; Weare 2004; Potter and Cess 2004; Zhang et al. 2005). Recently, more refined approaches stratifying the cloudiness and radiation data according to dynamical- or cloud-regimes have been published (Bony et al. 2004; Ringer and Allan 2004; Wyant et al. 2006; Jakob and Tselioudis 2003; Williams et al. 2005; Williams and Tselioudis 2007). We instead use a more regional approach. By analyzing the marine parts of the subtropics our purpose is to isolate the effects of marine low clouds in GCMs. The marine sub-tropics are influenced by the quasi-stationary highs associated with the subsiding leg of the Hadley circulation and vertical cloud growth is thus suppressed. In the eastern parts of the subtropical ocean basins, where upwelling is the cause of the relatively cold ocean water, large persistent decks of low marine stratiform clouds are found (e.g. Klein and Hartmann 1993). These clouds are mainly maintained by upside down convection when air at the top of the cloud becomes denser than underlying air due to radiative cooling and entrainment of dry overlying air. When the trade-winds advect the clouds equatorward the inversion-capped boundary layer deepens and there is a transition from stratocumulus to, less covering, trade-wind cumulus (see Fig. 1).

To successfully model low clouds and the transition between the different cloud types a good description of the vertical turbulence fluxes of heat, moisture and momentum is needed and rules for how these quantities are related to the cloudiness need to be known (Siebesma et al. 2003; Svensson et al. 2000).

Although there is a continuous development of faster computers which enables modelers to increase the resolution of GCMs, the resolution is still far too coarse to resolve individual clouds. Therefore clouds need to be parameterized from the prognostic grid-mean variables. Most of today's GCMs rely on two types of cloud parameterization—convective and stratiform parameterization. The former is often related to the updraft mass-flux, which depends on the stability of the atmospheric column, while the stratiform parameterization typically depends on the relative humidity of the grid-box.

The presence of low clouds significantly alters the otherwise low albedo of the ocean surface and such clouds thus have a large impact on the energy going into the ocean and hence on the climate system. Considering the large impact of marine low clouds on the Earth's radiation budget it is important that climate models manage to simulate their extent and physical characteristics. Systematic errors in the simulations of such clouds may limit the ability of the models to respond realistically to changes in antropogenic radiative forcing.

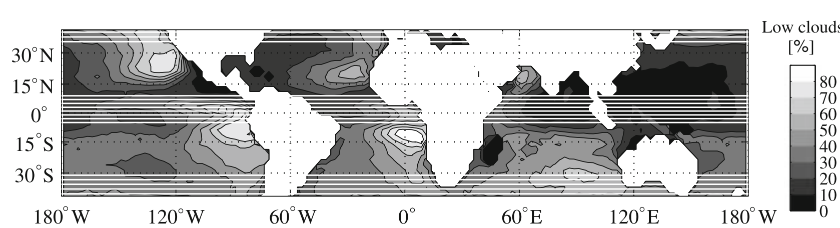


Fig. 1 ISCCP D2 mean-August low cloudiness for the period 1983–2001. The stripes indicate the area not included in the analysis

The fourth IPCC assessment report (IPCC, Solomon et al. 2007) identifies the response in cloudiness in a warmer climate as the primary source to the inter-model variability, and thus uncertainty, of future projections. Several earlier studies (Potter and Cess 2004; Webb et al. 2001; Weare 2004; Zhang et al. 2005; Ringer and Allan 2004) have reported the subtropics of GCMs to be associated with too small amounts of low level clouds, which in turn, are too optically thick. Compensation of errors enables the GCMs to still have a relatively good representation of the radiative fluxes at TOA. The purpose of this study is to investigate how the next generation of coupled ocean-atmosphere general circulation models, that contributed with simulations for the fourth IPCC assessment report (AR4), manage to simulates marine subtropical low level clouds for present climate. Here we examine their occurrence and radiative properties, and how this can be related to TOA radiative fluxes.

2 Data

In an effort to focus on marine low cloudiness in the models, we constrain our analysis to the ocean parts of the subtropics, thus avoiding the deep convection of the intertropical convergence zone (ITCZ) and the storm-tracks of the extratropics. The latitudinal bands considered are 5–30°S and 10–35°N (see Fig. 1), where the northward adjustment is due to the different distribution of land and ocean between the southern and northern hemisphere, resulting in a northward displacement of the ITCZ. A grid box is treated as marine when its land fraction is smaller than 20%.

2.1 Observations

We use monthly mean top-of-the-atmosphere observations of radiative fluxes from the earth radiation budget experiment (ERBE; Harrison et al. 1990) on a 2.5° by 2.5° grid. The radiative parameters included in the analysis are outgoing terrestrial and reflected solar radiation both for clear-sky and all-sky conditions. For the ERBE-measurement period (November 1984 to February 1990) monthly mean climatologies have been calculated. The monthly-mean clear-sky fluxes are missing for a couple of locations for a number of years, e.g. regions with persistent decks of low clouds. For these locations the climatological values have been calculated from the years when the monthly mean fluxes are available. This is of minor importance in the analysis since the number of grids that have missing clear-sky fluxes is small.

For cloudiness the monthly mean cloud-fraction climatologies from the international satellite cloud climatology project (ISCCP) D2 VIS/IR data (Rossow and Schiffer 1999), also presented on a 2.5° grid, have been used. The cloud fractions are divided into low, middle and high clouds based on their cloud-top pressure; low clouds having cloud tops below 680 hPa, middle clouds between 680 and 440 hPa and high clouds over 440 hPa. The period for which the ISCCP cloud parameter climatologies have been calculated is July 1983 to September 2001. As an example, Fig. 1 shows the August mean low cloudiness during the period, the persistent decks of low clouds in the eastern part of the subtropical ocean basins are clearly visible in the climatology.

In our analysis, we have given priority to climatologies calculated over a long time period rather than a complete temporal overlap of ERBE and ISCCP data. Since our focus is on the mean state of the data not on its variability or trends, this is justified.

For comparison with the models, monthly summaries of sea surface temperature (SST) from the international comprehensive ocean-atmosphere data set (ICOADS) have been used to calculate monthly mean climatologies for the period January 1980–December 1999.

2.2 Models

We have analyzed nine coupled atmosphere-ocean general circulation models (AOGCMs) that have contributed with simulations for the 4th Assessment Report of the IPCC. These nine models represent the subgroups of models which provide the monthly mean parameters needed for the analysis—a 3-D cloud distribution and TOA radiative fluxes—and which match the resolution of the gridded observation products. The simulation studied is the climate of the 20th Century experiment (20C3M) and model monthly-mean climatologies of the current parameters have been calculated over the period January 1980 to December 1999. Since the analyzed period is somewhat arbitrary, the sensitivity of this choice was tested by repeating the analysis with model monthly-mean climatologies calculated for the period November 1984 to February 1990 (the overlapping period where both ERBE and ISCCP data are available). The difference to this change was very small in all of the analyzed parameters and does not not change our conclusions.

The models included in the analysis are listed in Table 1. Most of the analyzed models have prognostic cloud condensate (CCSM3.0, MIROC3.2 [hires, medres], GFDL, and UK-HadCM3). In two of the models the large-scale cloudiness is treated prognostically (GFDL and UK-HadCM3) but in the other models the large-scale cloudiness is diagnosed from the relative humidity. The shallow and deep convection in the models are related to the

Table 1 The nine coupled models analyzed in this study

IPCC-AR4ID	ATM. Resolution	References
CCSM3	T85, L28	Collins et al. (2004)
FGOALS-g1.0	2.8° × 2.8°, L26	Yu et al. (2004)
GFDL-CM2.0	2.0° × 2.5°, L24	Delworth et al. (2006), Gnanadesikan et al. (2006)
GFDL-CM2.1	2.0° × 2.5°, L24	Delworth et al. (2006), Gnanadesikan et al. (2006)
IPSL-CM4	2.5° × 3.75°, L19	Marti et al. (2005)
MIROC3.2 (medres)	T42, L20	K-1 Model Developers (2004)
MIROC3.2 (hires)	T106, L56	K-1 Model Developers (2004)
PCM	T42, L26	Washington et al. (2000)
UKMO-HadCM3	2.5° × 3.75°, L19	Pope et al. (2000), Williams et al. (2001)

For more model information see http://www-pcmdi.llnl.gov/

vertical mass-flux. One model, CCSM3.0, has a specific parameterization for low marine clouds.

3 Method

Comparing satellite observations of cloudiness with model output is challenging. While the satellite, with its from-above perspective, only catches clouds not obscured by higher clouds, the models monthly mean cloudiness are defined on their vertical model levels. To make model data comparable to satellite observations, the different perspectives certainly have to be considered (Wei et al. 1996).

Probably the most consistent way to do this is by using the "ISCCP-simulator", which in the deduction of its satellite-perspective data, takes the properties of each model's radiation-scheme into account. Klein and Jakob (1999), Webb et al. (2001) and Zhang et al. (2005) are examples of studies employing the "simulator"-technique. The "simulator", however, needs to be run on-line with the model, and in our case, being served with monthly mean data, this is not an option.

Instead, we follow the method proposed by Weare (2004) that assumes maximum overlap for model layers contained in the ISCCP defined low, middle and high levels and then assuming random overlap of these three layers (see Fig. 2 for a schematic illustration). This procedure mimics the assumption of maximum-random cloud overlap implemented in many of today's GCMs. As discussed by Weare (2004), this method results in a smaller total cloud cover (just adding the low, medium high and high cloud fractions as seen from above) than what the models report. This contrasts to what Weare (1999) found when he applied the method to observations (surface observations were used to derive the three-dimensional structure of the clouds). In this case the method's reconstructed total cloud fraction overestimated the observed. An unavoidable disadvantage of processing the model layer output to a

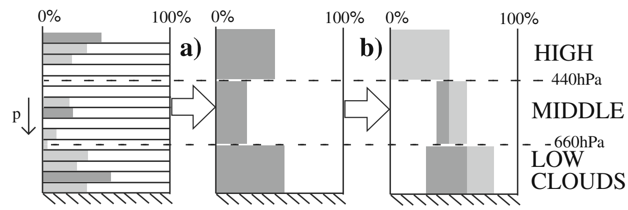


Fig. 2 Illustration of the process of transferring the data from model grid levels to ISCCP layers and finally to fractions of low and high level clouds. In the first step (**a**), the monthly-mean model layer cloud fractions are assumed to have a maximum overlap in the pre-defined ISCCP low, medium high and high layers. Secondly (**b**) these three layers are assumed to be randomly overlapped

satellite perspective is that the low cloudiness will be affected by possibly erroneous amounts of obscuring clouds.

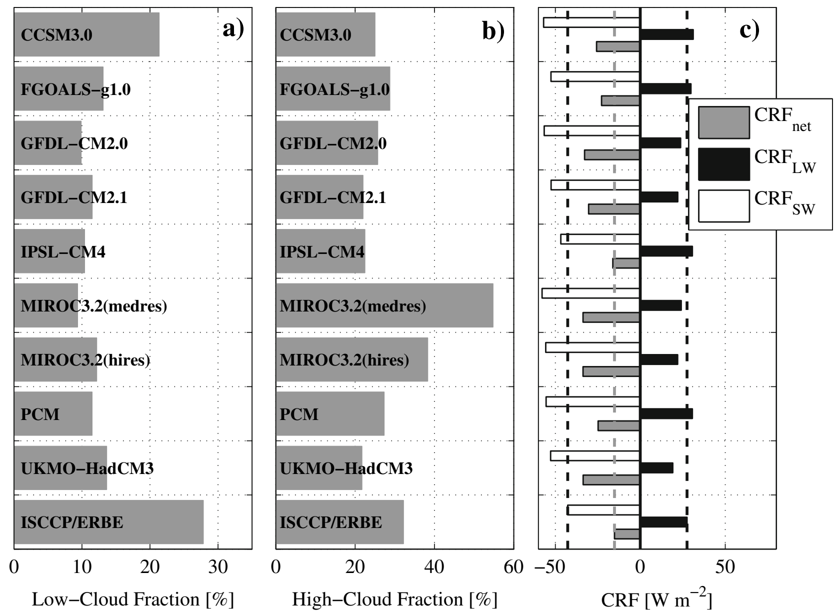
The relatively small amount of medium-high clouds and small inter-model difference of vertical layers found in the ISCCP-defined middle-cloud make it reasonable to include the middle (as seen from above) cloud fraction with the high level clouds. The combined medium-high and high cloud fraction will hereafter be referred to as high clouds.

As pointed out by Allan and Ringer (2003), different sampling processes of clear-sky fluxes between models and measurements result in inconsistencies in the cloud radiative forcing. These are however, mainly associated with regions of large-scale ascent (e.g. the ITCZ) where cloud-free regions are associated with substanially weaker ascent than the cloudy regions. Since the ascent is closely related to moisture, this results in measured clear-sky fluxes that systematically are from too "dry" satellite pixels.

4 Results

Model derived and observed climatological annual-mean low cloudiness for the marine subtropical region are summarized in Fig. 3a. All models underestimate the low cloud

Fig. 3 Marine subtropical model and observational climatological annual-mean **a** low cloudiness, **b** high cloudiness and **c** net cloud-radiative forcing (*gray*), longwave cloud-radiative forcing (*black*) and shortwave cloud-radiative forcing (*white*). The *dashed vertical lines* emphasize the observed values



fraction. This is in agreement with previous studies (Weare 2004; Zhang et al. 2005). The model mean bias in low clouds in the marine subtropics is –15.1%, which implies a model mean low cloudiness almost a factor two smaller than the observed. A possible reason why the CCSM3.0 model stands out, as being closer to the observations than the others, is the inclusion of a separate empirical parameterization (Slingo 1980; Klein and Hartmann 1993) of marine stratocumulus clouds. It should be emphasized that the amount of low clouds is "as seen from above", see Sect. 3. This implies that errors in high clouds propagate to the lower clouds, i.e. the underestimation of low clouds could be a consequence of too much high clouds. However, only the two versions of the MIROC-model have an excess of high clouds compared to the observations (Fig. 3b).

The net cloud-radiative forcing at the TOA (Fig. 3c) is a result of the whole cloud scene as well as the properties of the clear portion of the grid. It is surprising, keeping the underestimation of the low clouds in mind, that all models also show a too negative climatological annual-mean cloud-radiative forcing in this region. In other words, the models' net-CRF is more negative than observed by ERBE. The negative model mean net-CRF bias in the region is 13 W m^{-2}. It is obvious that the net-CRF model bias originates from too strong shortwave CRF in the models compared to the observations. For the longwave component of the cloud-radiative forcing, there is no obvious bias as for the shortwave component.

Similar global results, indicating that models overestimate the cloud radiative cooling, have also been reported in intercomparison studies by Potter and Cess (2004) and Weare (2004). Webb et al. (2001) found the trade-cumulus regions in models to be associated with an overestimation of the cloud radiative cooling while the regions dominated by stratocumulus instead underestimated the cloud radiative cooling. Both these cloud-regimes are included in our analysis domain.

As mentioned before the CRF is not only a result of the cloud scene but also depends on the clear-sky fluxes (see Eqs. 1 and 2). Outside the tropics, for example, there is a close relation between the outgoing clear-sky longwave radiation at TOA and the SST in such a way that a negative model bias in SST is expected to generate a negative bias in the clear-sky outgoing longwave radiation. In fact, all of the models included in the analysis except FGOALS, have a negative bias in the mean subtropical SST compared to ICOADS (model-mean bias –0.5 K) but there is no clear signal in the CRF_{LW} from this (Fig. 3c). It is possible that the negative bias in SST to some extent is a consequence of the too negative SW_{CRF} in the models. The significant correlations between models' SST bias and net-CRF bias (the correlations range between 0.34 to 0.64 for the models) suggest this could be the case.

All the models, except FGOALS, have a negative bias (compared to ERBE) in the reflected shortwave clear-sky flux, $SW\uparrow_{cs}$, indicating a too dark ocean surface or an overestimation of shortwave absorption in the atmosphere. However, it cannot be ruled out that the bias originates from an inconsistency of clear-sky sampling between models and observations. The model mean bias of $SW\uparrow_{cs}$ for the marine subtropics is –3.1 W m^{-2}, which only explains a small part of the model bias seen in the net-CRF (–13 W m^{-2}). The larger part of the net-CRF bias is coupled to the bias in the reflected shortwave all-sky flux ($SW\uparrow_{as}$), i.e. related to the cloudiness. The model mean bias in the $SW\uparrow_{as}$ is 8.4 W m^{-2}, suggesting too much reflection of clouds in the models. There is also a strong negative correlation of the net-CRF and the $SW\uparrow_{as}$ bias in the individual models. The mean model correlation coefficient between the net-CRF and $SW\uparrow_{as}$ biases is –0.76. This can be compared with the mean model bias correlation of net-CRF vs. $SW\uparrow_{cs}$, $LW\uparrow_{as}$ and $LW\uparrow_{cs}$, which are 0.19, 0.01 and 0.05, respectively.

Figure 4 shows distributions of net-CRF and low cloudiness for the nine models and the observations for all the grid-boxes confined in the marine subtropics. The net-CRF distributions have been normalized with the total number of gridpoints in such a way that a unit corresponds to one permil of the total amount of data.

It is evident from Fig. 4 that both models and observations have increasing amounts of low clouds associated with increasingly negative CRF. This is expected since the low clouds mainly affect the albedo and only have minor influence on the outgoing terrestrial radiation. There are however prominent differences in the distribution between most of the models and the observations. While observations indicate a moderate increase in cloud radiative cooling with increasing amounts of low clouds, models show a more pronounced increase. We want to re-emphasize that the CRF at the TOA is a product of all clouds not obscured by higher clouds and their individual radiation properties. Therefore there is no unambiguous relationship between the low cloud fraction and the CRF. This is apparent in several of the models where the smallest (and sometimes non-existing) amounts of low clouds are associated with positive net-CRF. This cloud radiative heating then has to be explained by higher clouds. It is also so, due to the cloud overlap, that the maximal low and high cloud fraction vary inversely, i.e. small amounts of low clouds allow large variations and amounts of the high cloud fraction. Correspondingly a large low-cloud fraction limits the amount and variation of the high-cloud fraction.

The corresponding distributions of net-CRF and high cloudiness are shown in Fig 5. The units (but not the gray scale) are comparable to the ones in Fig. 4. For the high clouds the distributions are more scattered and the scatter

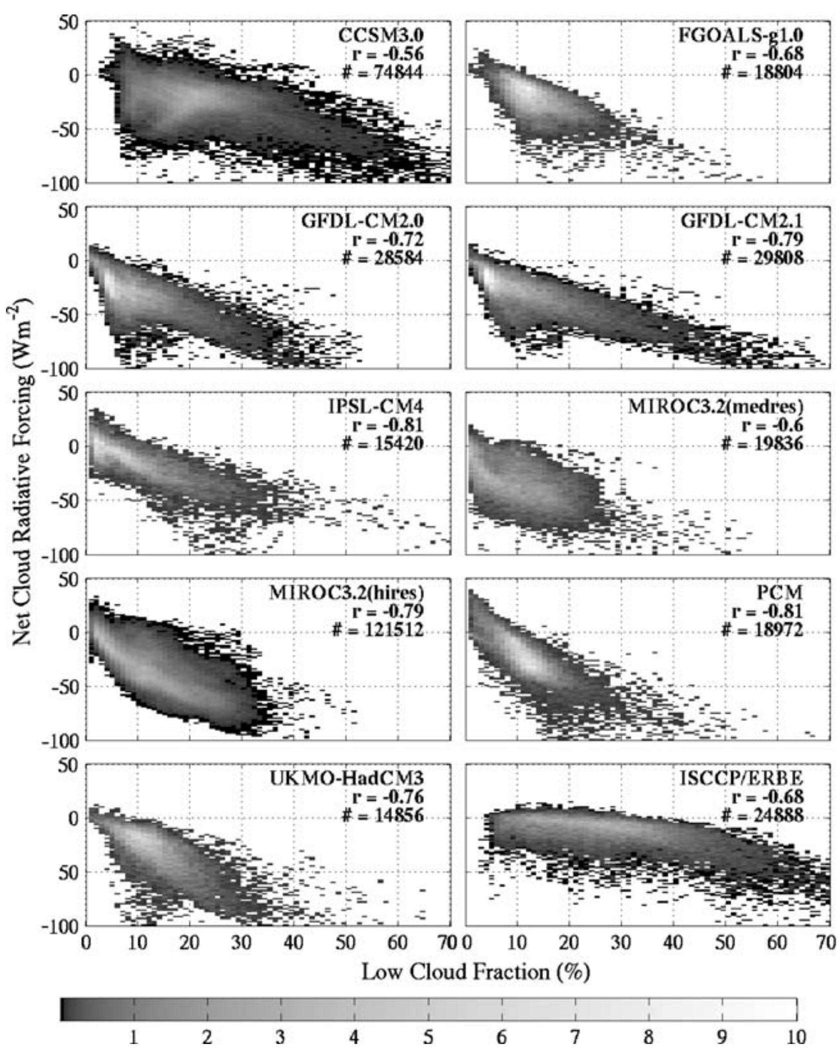


Fig. 4 Relation between climatological monthly mean low-cloud amount and CRF shown as density-scatter plots for the nine models and observations. Every grid-point confined in the subtropical marine region is represented by 12 samples—one for each month. The density-scatter bins have the size 1% × 1.5 W m^{-2} and the scale has been normalized such that unity represents one permil of the total samples. Correlation coefficients (*r*) and number of samples (#) are given in each panel

density does not reach the values it did for the low clouds. This is also visible in smaller correlation coefficients (Fig. 5). A weak trend of larger amounts of high clouds to be associated with less negative CRF can be seen in both models and observations.

The negative net-CRF associated with small high-cloud fractions is clearly due to low clouds and the before mentioned inverse relation between low and high cloud amounts. The largest variability in CRF in both models and observations is seen for small high-cloud fractions, where also the variability in low clouds is biggest (not shown).

5 Discussion and conclusions

As stated in the AR4 (IPCC, Solomon et al. 2007), realistic parameterizations of cloud processes are a prerequisite for reliable current and future climate simulations.

In this study radiative fluxes at TOA and 3D cloud fraction simulations from nine global coupled ocean-atmosphere climate models, participating in the IPCC AR4, have been analyzed and compared to ERBE and ISCCP data sets. Since our focus has been on the simulation of low clouds the analysis is confined to the marine subtropics.

One major conclusion is that all the analyzed GCMs underestimate the low level cloud amounts in the marine subtropics while they are overemphasizing cloud radiative cooling. This confirms what have been reported in earlier studies for slab-ocean models and earlier model-versions. Thus, in-spite of many efforts for improving the representation of these clouds, major problems remain (IPCC, Solomon et al. 2007).

The excess model cloud radiative cooling is mainly connected to the excess of model shortwave cloud reflectance. An additional, but smaller contribution to the overestimate of the cloud radiative cooling is an

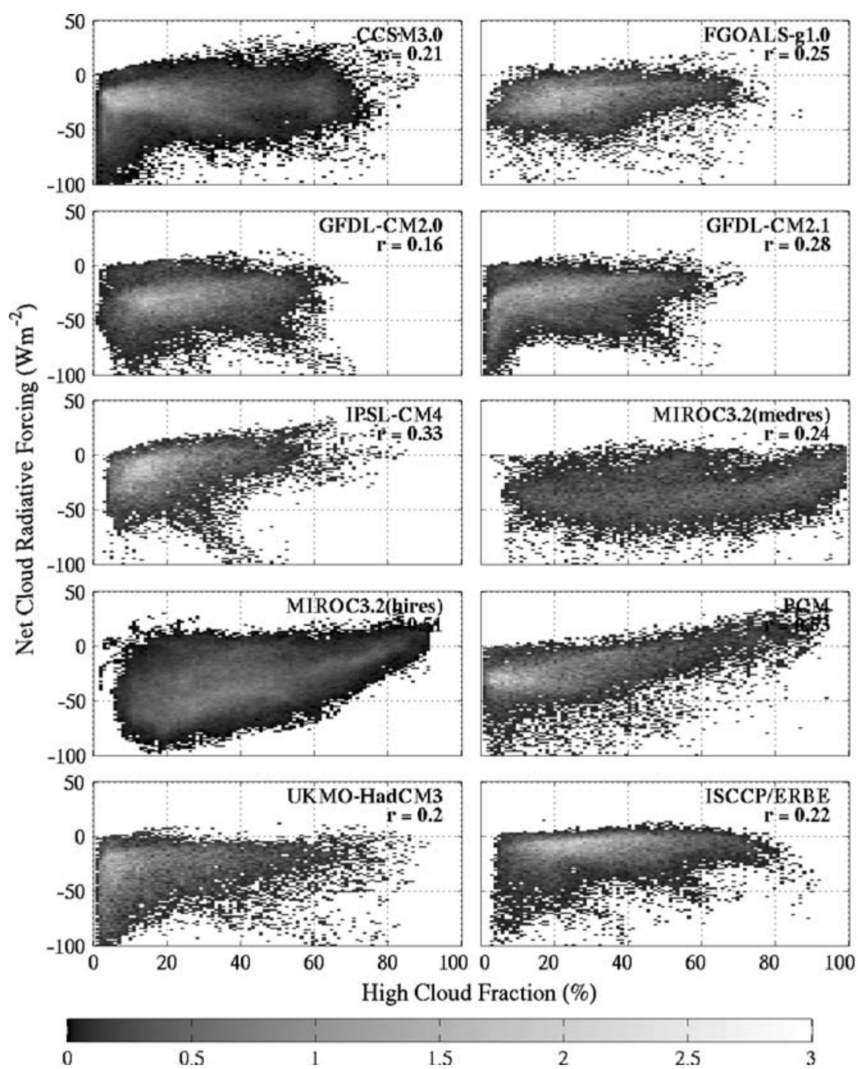


Fig. 5 Same as Fig. 4 (not the same scale) but now the relation between climatological monthly mean high-cloud amount and CRF

underestimate (compared to ERBE) of the shortwave clear-sky reflectance in the region. All models, except one, have a negative SST bias, which might be a consequence of the overestimated cloud radiative cooling.

Let us assume, as a first approximation, that the long-wave and shortwave contribution to the CRF from high clouds cancel each other. Then the mean CRF and the fractional area associated with a certain range of low cloudiness may be calculated. If we then fit polynomials to the CRF dependence on the low level cloud fraction (Fig. 4, we choose second order polynomials), it is possible to test the radiative response to a certain change in the low cloudiness, in the models as well as in the observational data (see Fig. 6). The change applied here is an absolute change in low level clouds (as seen from above), assuming the area distribution to remain unchanged, similarly to the early experiment by Slingo (1990). This is obviously a simplification of the complex response of the low clouds to a changing climate, but it serves to illustrate the sensitivity of the problem. Experiments of climate change either by a doubling of CO_2 (e.g. Bony and Dufresne 2005; Soden and Held 2006; Webb et al. 2006; Wyant et al. 2006) or a ±2K SST perturbation (e.g. Cess et al. 1990; Cess et al. 1996) do not give a clear signal of how the cloudiness will change.

In our analysis (Fig. 6), the combined data from ERBE and ISCCP indicate a 5% absolute increase in the low cloud cover to be associated with a decrease in the CRF of about 3 W m^{-2}. This value can be compared with a mean model decrease in CRF of about 10 W m^{-2} for the same cloud change. According to this method, it is obvious that models are more sensitive to a given change in cloudiness than what observations indicate. This is a direct consequence of the underestimation of the low cloudiness and overestimating of the cloud reflectance in the models compared to the observations. Since the area covered by marine low clouds in the subtropics make up 8% of the global surface, these biases may have global significance.

The results suggest that models have a too strong cloud feedback associated with a given, positive or negative, change in low cloudiness. A model that shows an increase in low cloudiness for a specific future scenario would then have a too strong negative cloud feedback and a model which show a decrease in the low cloudiness would then have a too strong positive cloud feedback compared with observations. This does not claim anything about the actual future change in low cloudiness, but suggests that a model, which correctly depicts the change in low cloudiness, will have a too strong radiative response.

Our results might seem to contradict the results of Bony and Dufresne (2005) who found that the sensitivity of interannual variations of net-CRF in regions associated with subsidence is smaller in AOGCMs than in observations (see their Fig. 3). However, this may not be a contradiction. In our analysis, we have assumed that modelled cloud changes are realistic. Further, we have implicitly assumed both modelled and observed clouds to maintain their optical characteristics during the hypothetical cloud change. Since Bony and Dufresne (2005) study AOGCMs none of these assumptions are made. Their reported smaller model net-CRF sensitivity in subsidence regions might therefore, for example, be due to smaller inter-annual variations of low cloud cover in the models than in the observations.

In a recent study on climate feedbacks in coupled climate models (Soden and Held 2006), the authors find a

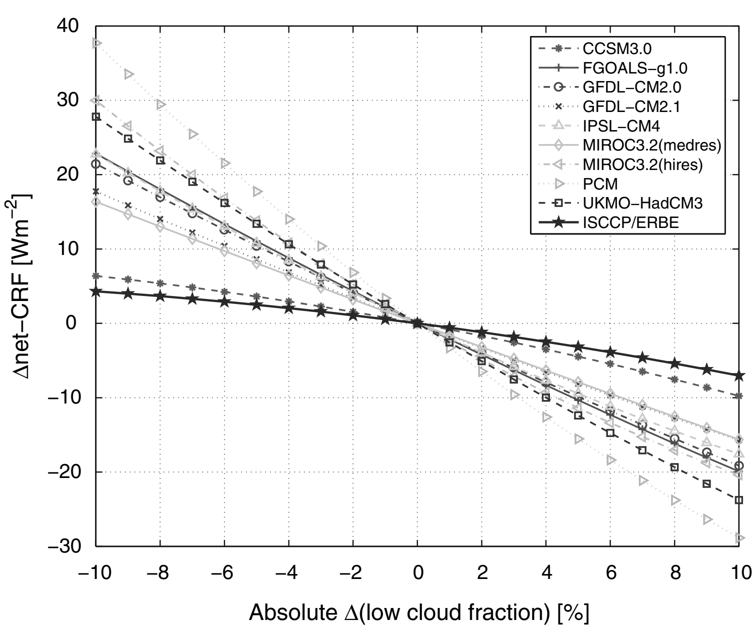


Fig. 6 Response in CRF to an hypothetical absolute change in the low cloud amount

positive cloud feedback in all of the analyzed models. Other studies have shown that the change in low cloudiness is the main contributor to the cloud feedback (e.g. Bony and Dufresne 2005). To the extent that our assumption about cloud overlap is realistic and that the spatial change in low cloudiness is more important than changes in optical thickness, our results together with the indications in these recent studies suggest that the treatment of low clouds in all investigated models tend to overestimate the climate sensitivities of the models due to the too strong cloud feedback.

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