

Climate response to the physiological impact of carbon dioxide on plants in the Met Office Unified Model HadCM3

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Abstract The concentration of carbon dioxide in the atmosphere acts to control the stomatal conductance of plants. There is observational and modelling evidence that an increase in the atmospheric concentration of CO₂ would suppress the evapotranspiration (ET) rate over land. This process is known as CO₂ physiological forcing and has been shown to induce changes in surface temperature and continental runoff. We analyse two transient climate simulations for the twenty-first century to isolate the climate response to the CO₂ physiological forcing. The land surface warming associated with the decreased ET rate is accompanied by an increase in the atmospheric lapse rate, an increase in specific humidity, but a decrease in relative humidity and stratiform cloud over land. We find that the water vapour feedback more than compensates for the decrease in latent heat flux over land as far as the budget of atmospheric water vapour is concerned. There is evidence that surface snow, water vapour and cloudiness respond to the CO₂ physiological forcing and all contribute to further warm the climate system. The climate response to the CO₂ physiological forcing has a quite different signature to that from the CO₂ radiative forcing, especially in terms of the changes in the temperature vertical profile and surface energy budget over land.

Keywords Carbon dioxide · Physiological forcing · Climate response · Feedbacks

1 Introduction

The increase in the atmospheric concentration of carbon dioxide (CO₂) has a direct effect on plant physiology through a decrease in the stomatal conductance, which decreases the evapotranspiration (ET) flux to the atmosphere (Collatz et al. 1992; Sellers et al. 1992). Change in the atmospheric CO₂ concentration is therefore expected to have a direct effect on climate, in addition to its indirect effect as a major greenhouse gas. Betts et al. (2007) refer to this effect as the “CO₂ physiological forcing”, and it has been shown to have an impact on the surface temperature both in laboratory experiments (Field et al. 1995), CO₂-enriched field experiment (Hungate et al. 2002; Long et al. 2006) and in climate model experiments (Sellers et al. 1996; Betts et al. 1997; Cox et al. 1999; Douville et al. 2000). The surface temperature is expected to increase in response to the reduced latent heat flux to the atmosphere so that the energy budget of the surface (composed of net shortwave radiation, net longwave radiation, latent heat and sensible heat fluxes) remains equilibrated. While this surface temperature increase is well observed in some models (e.g. Betts et al. 1997), it is not systematic in other models because of associated changes in soil wetness and atmospheric circulation (Douville et al. 2000).

Gedney et al. (2006) have shown that the spatial and temporal signature of the CO₂ physiological effect could be detected in the climate record of runoff. The modelled trend in runoff could only be made consistent with observations if the suppression of plant transpiration due to CO₂-induced stomatal closure was considered. Leipprand and Gerten (2006) have studied the response of ET, soil moisture and runoff under doubling atmospheric CO₂ concentration using an offline vegetation-hydrology model driven by climate model output. Gerten et al. (2007) then

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showed that ecosystems are less water limited than anticipated from changes in soil moisture alone because of the physiological vegetation response to elevated CO₂. Betts et al. (2007) further investigated the robustness of the runoff response due to the CO₂ stomatal conductance effect to model parametric uncertainties, using a climate model which included stomatal responses within its land surface scheme which hence allowed for land-atmosphere feedbacks. They found that the physiological effect of doubled carbon dioxide concentrations on plant transpiration increases simulated global mean runoff by 6% relative to pre-industrial levels. Douville et al. (2000) and Kergoat et al. (2002) argued that the CO₂ fertilisation effect on plants also contributes to increase the leaf area index (LAI), which compensates for the reduced stomatal conductance at the leaf level. However Cramer et al. (2001), Leipprand and Gerten (2006) and Betts et al. (2007) find that this offset is not total in most models. Likewise changes in vegetation dynamics due to increased concentration of carbon dioxide can also be responsible for a climate response. In particular Levis et al. (1999) and Bala et al. (2006) showed that CO₂ fertilization can also lead to an expansion of boreal forests which causes some warming because of the associated decrease in surface albedo.

It is well known that climate responds to the CO₂ radiative forcing with an increase in the evaporative flux over the ocean and an increase in atmospheric water vapour (e.g. Meehl et al. 2007). By changing the ET flux over land, the CO₂ physiological forcing can also potentially modify the water vapour content of the atmosphere. Since water vapour is the most important greenhouse gas in the atmosphere, there can be important feedbacks through the Earth's radiative balance. One may, *prima facie*, argue that a decrease in the ET flux should translate into a decrease in the water vapour column (or precipitable water), thus decreasing the greenhouse effect and contributing to a cooling effect. We will show this not to be the case because the atmospheric water vapour budget is as much controlled by sinks as it is by sources. In particular, conversion of water vapour into cloud water (which can then be removed by precipitation) requires the water vapour partial pressure to exceed the saturation water vapour pressure, which itself depends on the atmospheric temperature through the Clausius–Clapeyron relationship. This is why, to first approximation, the water vapour concentration in the atmosphere adjusts to changes in temperature in order to approximately maintain relative humidity constant. This has been a robust feature in climate models (Soden et al. 2002) and is also largely supported by observations of climate change (Trenberth et al. 2007). Since the CO₂ physiological effect not only affects the ET flux but also the surface temperature, the climate response of water vapour is not straightforward.

The purpose of this study is to examine the nature of the climate response to the CO₂ physiological forcing for a fixed distribution of vegetation type and LAI. Given the current uncertainties in the vegetation response to CO₂, it is appropriate to look at the climate response to increased LAI and CO₂ physiological forcing separately. While previous studies have looked at the response of surface temperature, soil moisture and runoff, we focus here on the response of the surface and the atmosphere from a thermodynamic point of view. In particular, we investigate how a forcing at the surface propagates to an atmospheric response and how it can alter the vertical structure of the atmosphere.

2 Experimental design

2.1 Model setup

We use the Met Office Hadley Centre HadCM3 coupled ocean-atmosphere model for this study (Gordon et al. 2000). This version of HadCM3 includes the MOSES land surface scheme (Cox et al. 1999) which models the plant physiological processes of photosynthesis, respiration and transpiration, enabling plant responses to changes in meteorological conditions and atmospheric CO₂ to affect the surface water budget as observed in experimental studies. An increase in CO₂ concentration at the leaf surface results in a reduction in stomatal conductance to water vapour, and hence a decrease in transpiration. Stomatal conductance also depends on temperature, humidity, soil moisture content, photosynthetically active radiation and LAI, with the aggregate effect scaled with LAI being termed canopy conductance.

Net leaf photosynthesis, stomatal conductance and CO₂ concentration within the stomatal cavities are related to each other and to leaf surface temperature and external CO₂ concentration through the closure suggested by Jacobs (1994), which further depends on the humidity deficit at the leaf surface. Photosynthesis is simulated following Collatz et al. (1991) and Collatz et al. (1992) for C3 and C4 plants, respectively, but with an additional dependence on soil moisture from Cox et al. (1999). Scaling of the physiological processes from the leaf to the canopy scales is carried out assuming that the factors limiting photosynthesis [mean incident photosynthetically active radiation (PAR), leaf nitrogen content and maximum rate of carboxylation of RuBisCO] vary proportionally within the canopy (Sellers et al. 1992). The overall canopy conductance decreases with canopy humidity deficit, increases with PAR, soil moisture availability and LAI, and peaks at an optimum canopy temperature. It also peaks at an optimum canopy CO₂ concentration of approximately 0.15 g kg⁻¹ for C3 and 0.02 g kg⁻¹ for C4 plants (see Fig. 1).

The model is run for the period 1860–2100 under the IS92a emission scenario. We use a pair of experiments which have fixed vegetation cover and LAI so that the CO₂ physiological effect can be better isolated. We consider two transient climate experiments, where the control (RAD) and perturbed (RADPHYS) experiments have the CO₂ physiological effect switched off and on, respectively. This means that in RAD, the plant physiology calculations are carried out with CO₂ concentration fixed at the initial concentration throughout the simulation, while those in RADPHYS are carried out with time-varying CO₂ concentration as used in the radiation budget calculations in both simulations. Note that the default version of HadCM3 corresponds to RADPHYS as it includes the CO₂ physiological effect. The two simulations start from the same initial conditions of climate and CO₂ concentration at 1860.

2.2 Result analysis

For each experiment EXP (either RAD or RADPHYS), we diagnose the change in climate parameter X between the last 30 years of the twentieth century and the last 30 years of the twenty-first century:

$$\Delta X_{\text{EXP}} = X_{\text{EXP}}(2070\text{--}2100) - X_{\text{EXP}}(1970\text{--}2000)$$

We then define the climate response to the CO₂ physiological effect as the difference in ΔX between the RADPHYS and the RAD experiments:

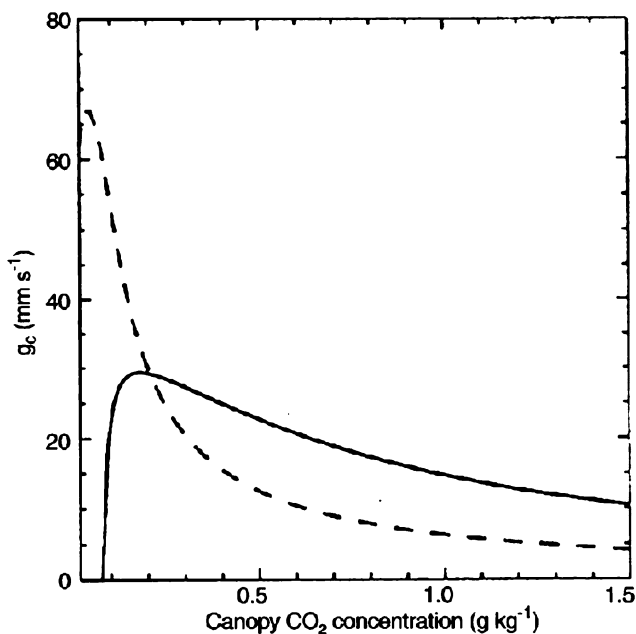


Fig. 1 Variation of canopy conductance, g_c , with canopy CO₂ concentration in MOSES-2, with other environmental influences held constant, for C3 (solid line) and C4 (dashed line) grasses (after Cox et al. 1999)

$$\Delta X = \Delta X_{\text{RADPHYS}} - \Delta X_{\text{RAD}}$$

By adopting this methodology the effects of the CO₂ physiological forcing are diagnosed under a future, warmer climate where the patterns of water stress and the CO₂ physiological forcing are different from that in the present-day climate. We will also provide some information on which changes are statistically significant. For this purpose the annual means have been detrended for each of the 30-year periods in each experiment. Means and standard deviations have then been calculated and Monte-Carlo simulations with 10^6 members were performed on the double difference ΔX . We then test if the differences in RAD and RADPHYS variables are significant at the 5% confidence level. Note that the data have only been detrended to assess statistical significance, with the data in Table 1 itself not being detrended.

As we will see in the next section, the land surface warms more in the RADPHYS than in the RAD experiment. Some of the differences observed between the RADPHYS and RAD experiments may just be a response to this additional surface warming rather than a specific response to the CO₂ physiological forcing. We will therefore investigate whether the strength and pattern of the climate feedbacks in response to the CO₂ physiological forcing are different to those in response to the CO₂ radiative forcing. To do so we normalise ΔX_{RAD} and ΔX by the corresponding land surface warming and compare the two normalised responses:

$$\Delta X_{\text{RAD}_N} = \Delta X_{\text{RAD}} / \langle \Delta \text{TS}_{\text{RAD}} \rangle$$

$$\Delta X_N = \Delta X / \langle \Delta \text{TS} \rangle$$

where TS is the surface temperature and $\langle \rangle$ denotes a global land average. We found this to be a meaningful approach because the variables we have examined respond almost linearly with surface temperature change.

3 Results

3.1 Surface temperature

Our experiments confirm the fact the CO₂ physiological effect is responsible for an additional surface warming due to the suppressed evaporative cooling at the land surface. This extra warming amounts to 0.52 and 0.14 K over land and ocean, respectively. The changes in surface temperature between the RAD and RADPHYS experiments as summarized in Table 1 have been found to be statistically significant at the 5% level. The warming over land is concentrated at mid and high latitudes of the northern hemisphere but is also present over other land areas and in particular over South America (Fig. 2). The land-sea

Table 1 Changes in climate variables between the last three decades of the twentieth century and the last three decades of the twenty-first century for the RAD and RADPHYS experiments and their difference

Twenty-first century changes in	RAD	RADPHYS	RADPHYS–RAD
Globe 1.5 m surface temperature (K)	+2.55	+2.80	+0.25
Land 1.5 m surface temperature (K)	+3.51	+4.03	+0.52
Ocean 1.5 m surface temperature (K)	+2.16	+2.30	+0.14
Land-sea contrast (K)	+1.35	+1.73	+0.38
Net SW surface radiation—global average (W m ⁻²)	-1.33	-0.89	+0.43
Net LW surface radiation—global average (W m ⁻²)	+4.54	+4.55	+0.01
Sensible heat flux—global average (W m ⁻²)	-0.26	+0.20	+0.46
Latent heat flux—global average (W m ⁻²)	+2.69	+2.56	-0.13
Net SW surface radiation over land (W m ⁻²)	+1.20	+2.85	+1.65
Net LW surface radiation over land (W m ⁻²)	+1.43	+0.48	-0.95
Sensible heat flux over land (W m ⁻²)	+3.43	+5.35	+1.91
Latent heat flux over land (W m ⁻²)	-0.81	-2.04	-1.23
Downward LW surface radiation over land (W m ⁻²)	+20.36	+22.48	+2.12
Upward LW surface radiation over land (W m ⁻²)	+18.94	+22.01	-3.07
Runoff over land (mm day ⁻¹)	+0.08	+0.11	+0.03
Precipitable water—global average (kg m ⁻²)	+4.79	+5.23	+0.44
Precipitable water—land average (kg m ⁻²)	+3.90	+4.36	+0.46
Total cloud cover (%)	-0.30	-0.56	-0.26
Total cloud cover over land (%)	-1.33	-2.25	-0.92
Planetary albedo (%)	-0.42	-0.61	-0.19
Total precipitation over land (mm day ⁻¹)	+0.05	+0.04	-0.01

The difference (last column) isolates the change in climate variables due to the CO₂ physiological forcing. The net radiative fluxes are from the atmosphere to the surface, therefore a positive net radiative flux indicates a net energy gain for the surface. All the changes in the last column are statistically significant at the 5% confidence level (using the detrended data) except for the change in land precipitation

contrast in temperature (defined as the difference in global land and ocean temperatures) increases by 1.35 and 1.73 K over the twenty-first century in the RAD and RADPHYS experiments, respectively. Therefore a 0.38 K increase in the land sea contrast is attributable to the CO₂ physiological effect. The land-sea ratio in surface warming (defined as the average warming over land divided by the average warming over sea over the period 1860–2100) is 1.65 and 1.73 in the RAD and RADPHYS experiments, respectively.

3.2 Land surface energy budget

Both the CO₂ radiative and physiological forcings cause significant changes in the global and land surface energy budgets (Table 1). We observe a large increase of 4.54 W m⁻² in the net downward LW radiation in response to the CO₂ radiative forcing, which is accompanied by a large increase in the latent heat flux of 2.69 W m⁻². The global-mean changes in RADPHYS are clearly dominated by the oceanic response to the CO₂ radiative forcing. However, over land, the response to the physiological forcing is of similar magnitude to the response to the

radiative forcing. In particular the reduction in the latent heat flux is 2.5 times larger in RADPHYS than in RAD. The CO₂ physiological forcing reduces the latent heat flux to the atmosphere by 1.23 W m⁻² by the end of the century. This is more than compensated by an increase in the sensible heat flux to the atmosphere (see Table 1) because there is also a net gain in radiative energy for the surface despite some compensating effects between the SW and LW radiation. The net SW flux to the surface increases by 1.65 W m⁻² due to an increase in the transparency of the atmosphere—mostly as a result of a reduced low cloud fraction—and a decrease in surface albedo. The net LW flux to the atmosphere increases because the surface upward LW flux increases more than the surface downward component (3.07 and 2.12 W m⁻², respectively). This is a result of increased surface temperature and increased greenhouse effect (discussed below). Overall the net increase in SW flux to the surface is larger than the net decrease in LW flux to the surface, leading to an increase in net radiative energy to the surface and a large increase in the sensible heat flux to the atmosphere (1.91 W m⁻²).

The CO₂ physiological forcing causes a decrease in surface albedo because of a reduction in the snow cover in

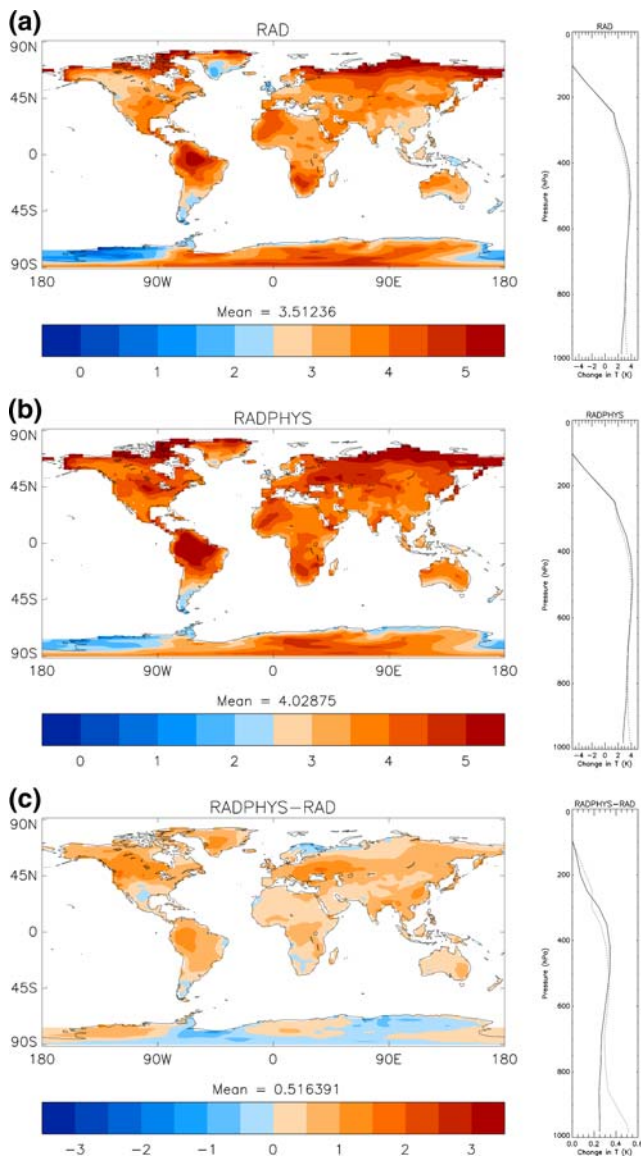


Fig. 2 Change in the surface (1.5 m) temperature (K) over land between the last three decades of the twentieth century and the last three decades of the twenty-first century **a** in the RAD experiment, **b** in the RADPHYS experiment, and **c** due to the CO₂ physiological effect (RADPHYS–RAD). The right panels show the corresponding change in the vertical profile of the globally averaged (*solid line*) and land-averaged (*dotted line*) temperature (K)

boreal regions. The surface albedo decreases by 0.0018 on average over land, but the decrease can be as large as 0.02–0.03 in boreal regions. We will discuss the surface albedo response further in Sect. 4.1.

The CO₂ physiological forcing not only increases continental runoff in the future but it also increases the soil moisture content. It turns the change in soil moisture from a negative trend (in RAD) to a positive one (in RADPHYS). The increase occurs over most of the continents

and has been shown to be a robust response to the CO₂ physiological forcing in a set of perturbed-physics experiments (Betts et al. 2007).

3.3 Land precipitation and runoff

Both RAD and RADPHYS simulate an increase in runoff throughout the twenty-first century. At 0.03 mm day⁻¹ the increase in runoff between RAD and RADPHYS is statistically significant. This magnitude of this increase is comparable to the increase reported by Betts et al. (2007) using HadSM3—a version of HadCM3 with a mixed layer ocean model—simulations. Both RAD and RADPHYS simulate a very small increase in land precipitation throughout the twenty-first century (0.05 and 0.04 mm day⁻¹, respectively, equivalent to 2.4 and 1.7% of the land precipitation), but the difference between the two simulations is small. The small decrease in RADPHYS as compared to RAD might be expected as a result of reduced transpiration in RADPHYS, with less moisture being “recycled” to the atmosphere in continental interiors. However, the change between the two simulations is not statistically significant. In contrast, Betts et al. (2007) found that, in two large ensembles of HadSM3, doubling CO₂ led to a greater precipitation increase on average in the ensemble which included physiological forcing compared to the ensemble which did not. The ensemble with physiological forcing already had higher precipitation in the control runs so the average percentage increase in precipitation was the same in both ensembles. Betts et al. (2007) suggested that the influence of reduced water recycling on precipitation was offset by increased moisture convergence due to higher surface temperatures with physiological forcing (due to reduced transpiration). However they did not sample the other parametric uncertainties equally in their two sets of experiments. Therefore it is difficult to conclude on this apparent discrepancy between the HadCM3 and HadSM3 models.

3.4 Atmospheric water vapour

As expected, atmospheric water vapour increases in both the RAD and RADPHYS experiments (Fig. 3). The CO₂ physiological forcing causes a further increase in atmospheric water vapour that is concentrated in the lower levels. It is interesting that the increase in precipitable water due to the CO₂ physiological forcing is about the same over land and ocean (Table 1) despite the fact that land warms more. This suggests that the water vapour change does not quite follow the Clausius–Clapeyron relationship over land in the RADPHYS experiment. These results are interpreted further in Sect. 4.2.

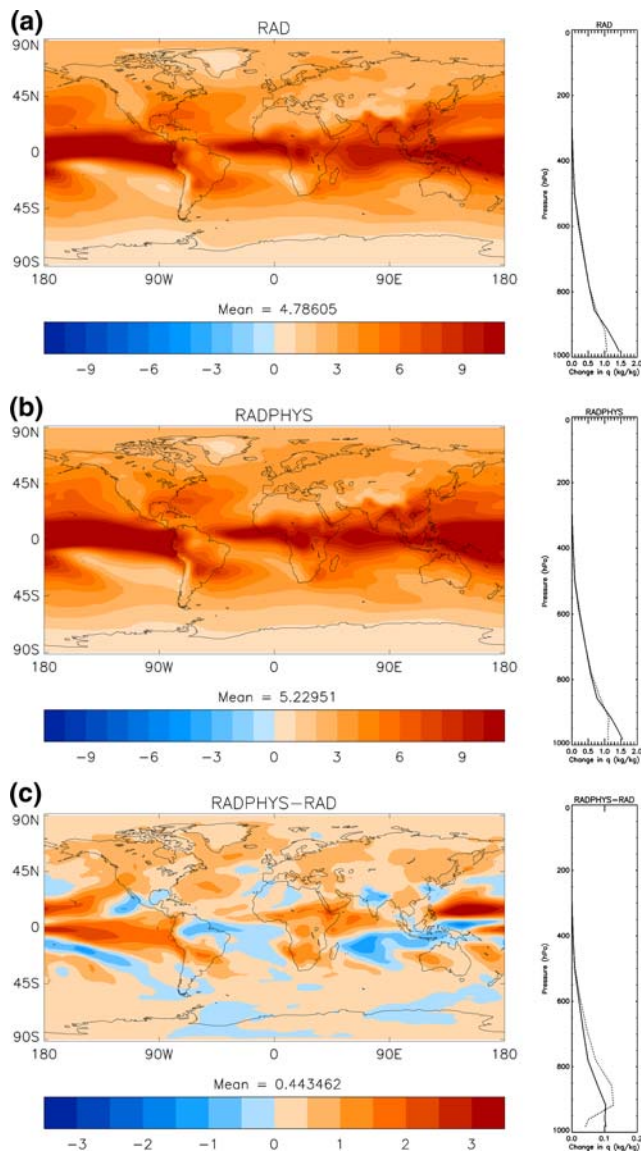


Fig. 3 Same as Fig. 2 but for the change in precipitable water (kg m^{-2}). The right panels show the corresponding change in the vertical profile of the globally averaged (*solid line*) and land-averaged (*dashed line*) water vapour content (kg kg^{-1})

4 Discussion

4.1 Surface feedbacks

We now investigate further the response of the atmosphere over land and compare the climate response to the CO₂ physiological forcing to that due to the CO₂ radiative forcing. To this end we normalise the climate response by the globally averaged land surface temperature change as described in Sect. 2. The normalised land surface temperature response is much more heterogeneous for the physiological than for the radiative forcing (Fig. 4) with some continental regions exhibiting twice as much surface

warming than the land average. Some of these regions appear to coincide with regions of large snow albedo change (Fig. 5). The snow albedo response is approximated here as the percentage change in surface albedo over land (not weighted by the amount of incoming solar radiation at the surface). The snow albedo response is almost twice as strong in response to the CO₂ physiological forcing ($-0.36\% \text{ K}^{-1}$) than in response to the CO₂ radiative forcing ($-0.21\% \text{ K}^{-1}$). It increases approximately linearly with surface temperature in both the RAD and RADPHYS experiments. This linear behaviour has also been found by Hall (2004) in a different climate model. The slope of the temperature-snow albedo relationship is slightly larger in the RADPHYS than in the RAD experiment, which explains the larger efficiency of the snow albedo response in RADPHYS. This is likely to be due to the heterogeneous nature of the temperature response which is larger than average in regions with snow cover. However it is difficult to ascertain whether the physiological forcing triggers a strong snow albedo feedback locally. This would require control experiments where the surface albedo is held fixed, but such experiments are difficult to design properly. Instead we look at how the change in surface temperature in boreal regions varies with the amount of snow present at the start of the run. A positive correlation between $\Delta\text{TS_RAD}$ or $\Delta\text{TS_RADPHYS}$ and the snow amount is an indication that there is a local snow feedback (Fig. 6a, b). However ΔTS itself is negatively correlated with the snow amount (Fig. 6c). This indicates that the temperature response to the CO₂ physiological forcing is less in regions with larger snow amount and longer snow season; this effect is stronger than any local potential snow albedo feedback. So there is no evidence that the snow albedo feedback is responsible for the local maximum in temperature response to the CO₂ physiological forcing observed in boreal regions.

4.2 Atmospheric feedbacks

The patterns of the normalised zonally averaged atmospheric temperature change over land are quite different for the two forcings (Fig. 7). There is an amplification of the surface temperature warming in the tropical troposphere in response to the CO₂ radiative forcing which has been a robust feature in climate models and some observational datasets (Santer et al. 2005). In contrast, the temperature response is a maximum at the surface and decreases with altitude for the physiological forcing, thus increasing the lapse rate. A secondary maximum in the upper tropical troposphere is also observed. Specific humidity also responds in a different manner, with a consistent increase throughout the troposphere for the radiative forcing (following the Clausius–Clapeyron law to first order) and a

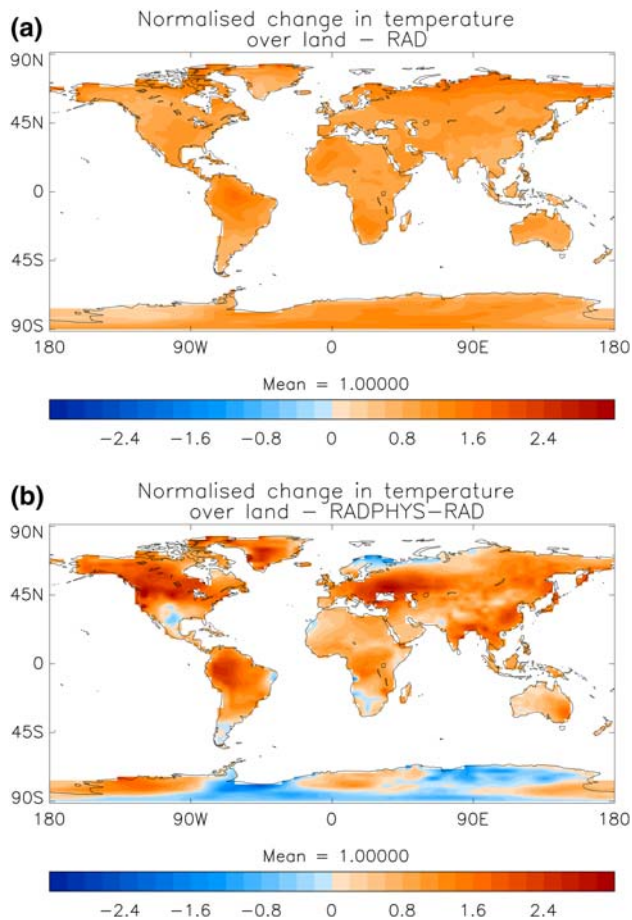


Fig. 4 Change in the surface (1.5 m) temperature over land normalised by the globally averaged surface temperature increase (unitless) for **a** the CO₂ radiative forcing and **b** the CO₂ physiological forcing

general but patchy increase for the physiological forcing (Fig. 8). The change in specific humidity is associated with a small decrease in relative humidity in the case of the CO₂ radiative forcing (Fig. 9a). There is an even larger decrease in relative humidity in response to the CO₂ physiological forcing but it is concentrated in the lower troposphere and only partly mirrors the change in specific humidity (Fig. 9b). This demonstrates that the water vapour content change does not quite follow the Clausius–Clapeyron relationship over land (as relative humidity is reduced). Moreover the increase in specific humidity despite a decrease in the latent heat flux over land suggests that the water vapour content is controlled by the sink as much as by the source term. Some of the increase in water vapour will be driven by the increase in latent heat flux over the ocean but it is important to note that the atmospheric reservoir (i.e. water vapour content) can increase even though the input flux to the atmosphere (i.e. evaporation) decreases. The decrease in specific humidity over the equatorial lower troposphere is associated with a decrease in evaporative flux in the tropical vegetation. Finally, the changes in

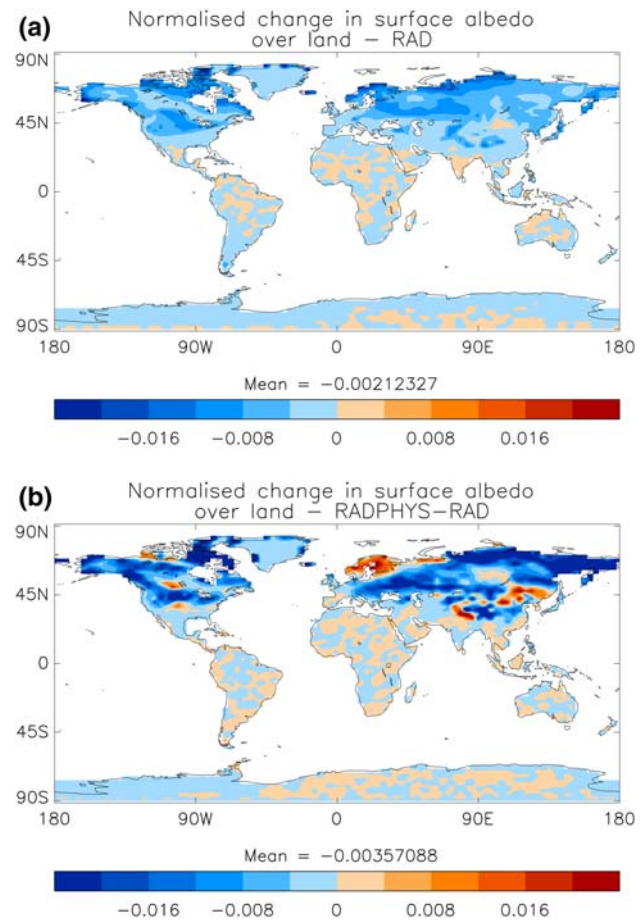


Fig. 5 Same as Fig. 4 but for the normalised change in surface albedo (K⁻¹)

layer (i.e. stratiform) cloud closely follow the respective distributions of the change in relative humidity (Fig. 10) as expected from the model parametrisation for layer cloud.

4.3 Comparison to other forcings of climate change

The global and land-averaged changes in latent heat flux of -0.13 and -1.23 W m⁻² over the twenty-first century correspond to reductions in the evaporation flux of -950 and $-2,500$ km³ year⁻¹, respectively. Since the CO₂ physiological forcing is not the only anthropogenic forcing mechanism that is expected to change evaporation at the Earth's surface, it is useful to compare the magnitude of the change in evaporation flux due to the CO₂ physiological forcing with that from other forcings:

1. As discussed above the evaporation flux responds to the CO₂ radiative forcing. The global increase in latent heat flux of 2.69 W m⁻² over the twenty-first century corresponds to an increase in the evaporative flux of 0.1 mm day⁻¹ or $19,000$ km³ year⁻¹. However, in HadCM3, this increase occurs mostly over the ocean.

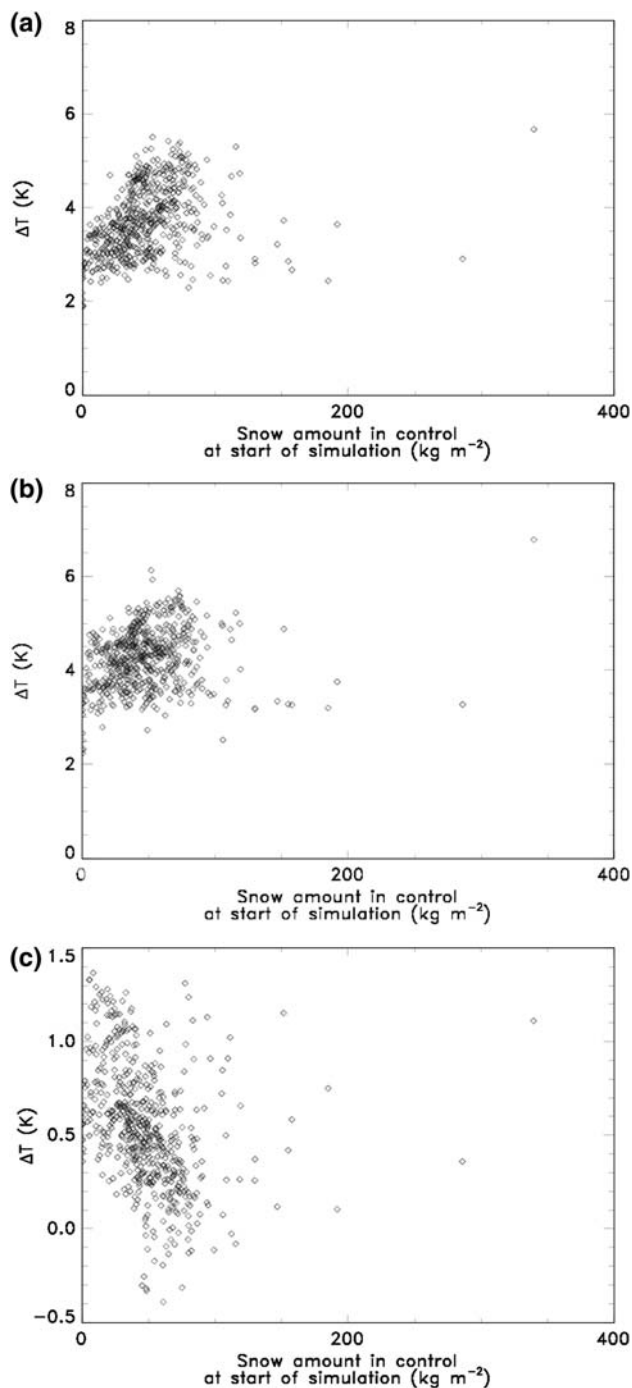


Fig. 6 Change in surface temperature (K) as a function of the snow amount at the start of the simulation (kg m^{-2}) for **a** the RAD experiment and **b** the RADPHYS experiment. **c** Change in surface temperature due to the CO₂ physiological effect (RADPHYS–RAD) as a function of the snow amount at the start of the RAD experiment. Only land data points north to 45°N excluding Greenland are considered

2. Anthropogenic aerosols scatter and absorb solar radiation, which results in a radiative forcing at the top of the atmosphere which can be either negative or

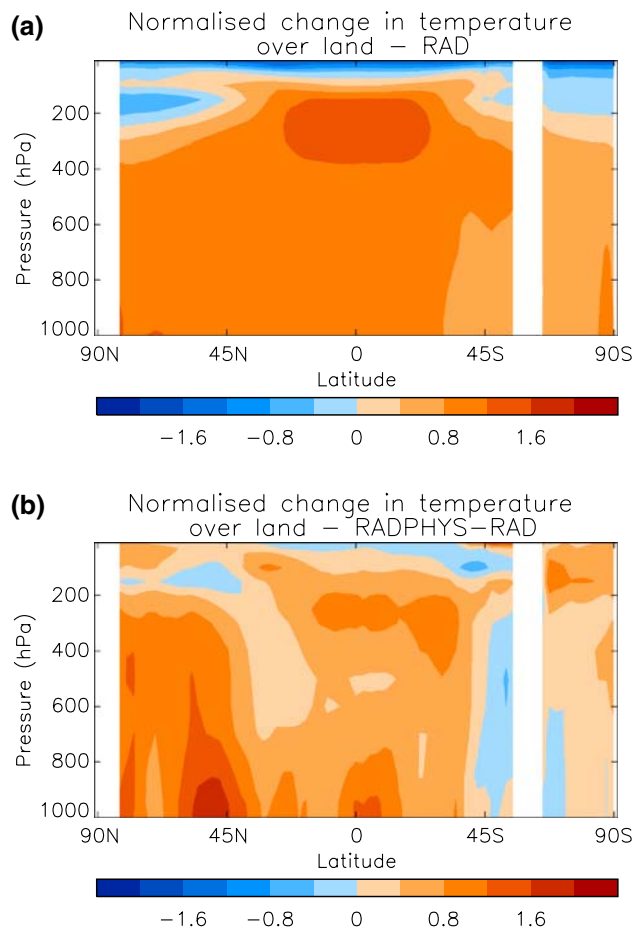


Fig. 7 Same as Fig. 4 but for the normalised change in zonally averaged temperature over land (unitless)

positive depending on the aerosol single scattering albedo, backscatter fraction, and surface albedo (Haywood and Boucher 2000). Anthropogenic aerosols also act as cloud condensation nuclei, thereby brightening clouds and affecting their precipitation efficiency. The aerosol direct and indirect effects result in a radiative forcing that is negative at the surface. This reduction in the amount of solar radiation available at the surface can be substantial on the regional scale with some impact on the hydrological cycle (Ramanathan et al. 2001). Using a climate model, Liepert et al. (2004) showed that reductions in surface solar radiation due to aerosols lead to weaker latent and sensible heat fluxes and hence to reductions in evaporation and precipitation despite global warming.

3. Both evaporation of water consumed for domestic and industrial use (e.g. from power plants) and emission of water vapour from fossil fuel or biomass combustion at the ground and from aviation are regarded as negligible sources of atmospheric water vapour as compared to natural sources. However, changes in evaporation

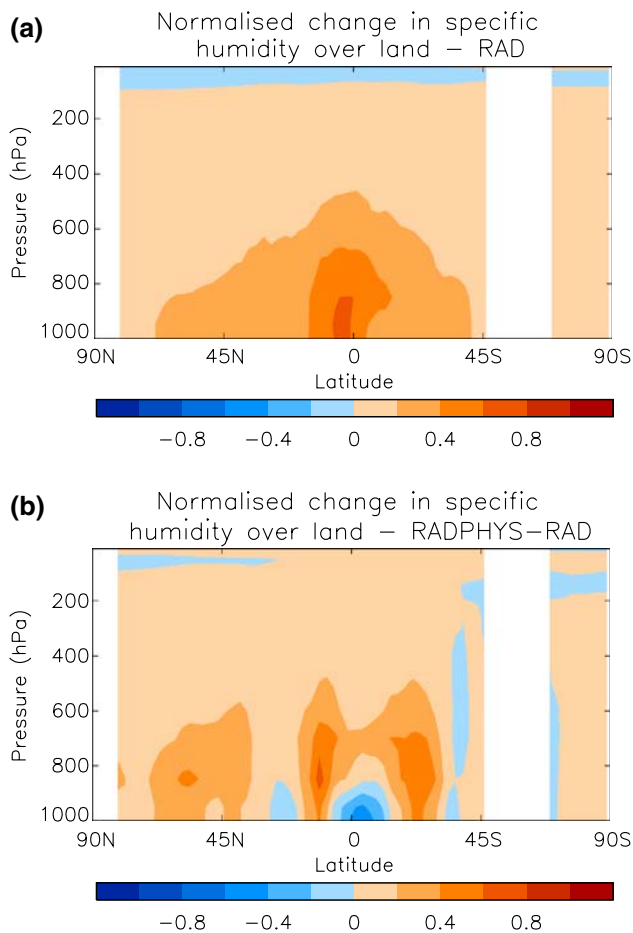


Fig. 8 Same as Fig. 4 but for the normalised change in zonally averaged water vapour concentration over land ($\text{kg kg}^{-1} \text{K}^{-1}$)

due to changes in land use and evaporation induced by irrigation can be important in some regions (Boucher et al. 2004, hereafter BMM04; Gordon et al. 2005; Lobell et al. 2006). Deforestation is estimated to have decreased global vapour flows from land by $3,000 \text{ km}^3 \text{ year}^{-1}$ or 4% (Gordon et al. 2005). The estimate of increased evaporation due to irrigation ranges from about $1,000 \text{ km}^3 \text{ year}^{-1}$ (BMM04) to about $2,000 \text{ km}^3 \text{ year}^{-1}$ (Milly and Dunne 1994) and $2,600 \text{ km}^3 \text{ year}^{-1}$ (Gordon et al. 2005). Change in ET has also been put forward by Bala et al. (2007) to explain the climate response to deforestation in their model simulations.

4.4 Conceptual model

Since the changes in the evaporation flux are of the same magnitude over land for the (present-day) irrigation and the (future) CO₂ physiological forcings, it is appropriate to discuss their relative effect in more details. We do this in the context of a conceptual model, which we first use to

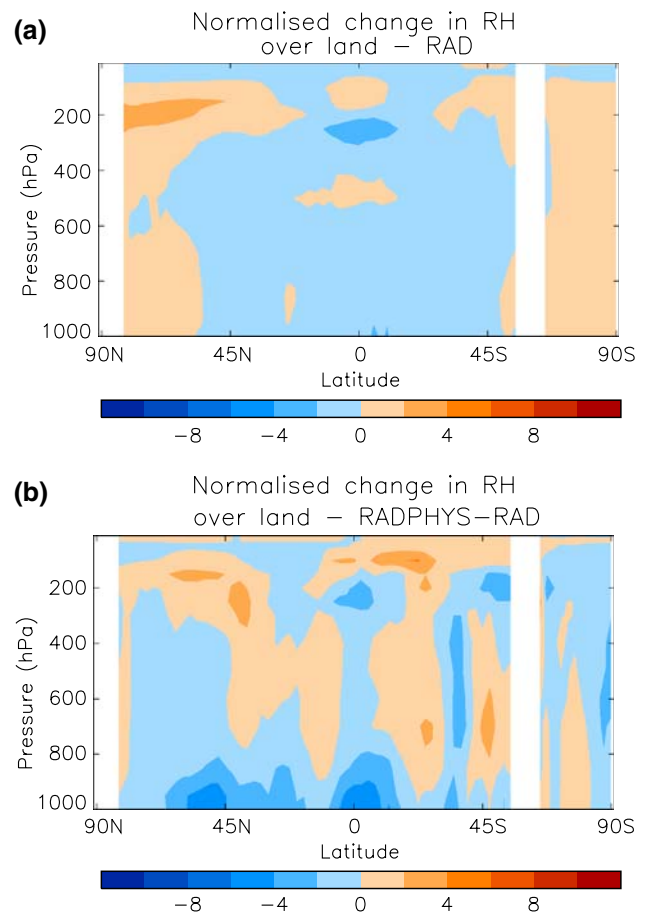


Fig. 9 Same as Fig. 4 but for the normalised change in zonally averaged relative humidity over land ($\% \text{K}^{-1}$)

explain the atmospheric temperature response to the CO₂ physiological forcing. A suppression of the ET flux at the surface will increase the lapse rate, moving away from the wet adiabatic rate towards the dry adiabatic rate. If we first assume that there is no associated changes either in the Earth's albedo or in the greenhouse effect, the change in lapse rate has to occur around a neutral point which corresponds to the average level where outgoing longwave radiation (OLR) originates in order for the energy balance to be maintained (Fig. 11a). Note that Joshi et al. (2007) similarly invoke the limitation of land evaporation to explain the land-sea contrast in warming. Put differently, the difference in climate response over land with and without the CO₂ physiological forcing resembles the contrast in response between land and ocean for the CO₂ radiative forcing. In summary, and neglecting for now changes in surface albedo, atmospheric water vapour and cloudiness, our conceptual model suggests that the suppression of the ET flux will result in an increase in the lapse rate with a warming in the lower troposphere and a cooling in the upper troposphere. We do indeed observe an increase in the lapse rate and a warming in the lower

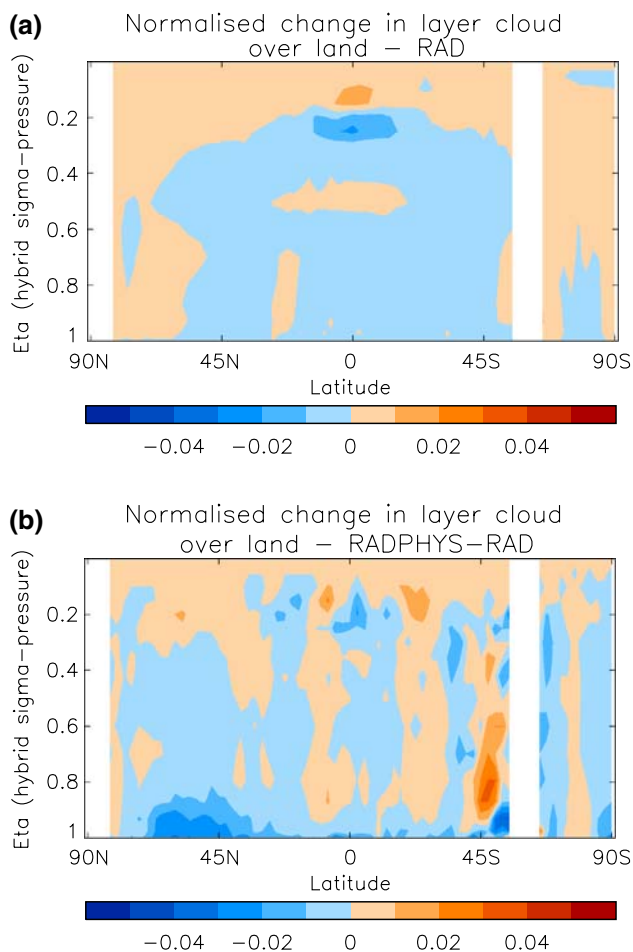


Fig. 10 Same as Fig. 4 but for the normalised change in zonally averaged layer cloud over land (K^{-1})

troposphere over land in our experiments; however, there is no cooling in the upper troposphere (see bottom right plot in Fig. 2).

As discussed in the introduction, the water vapour content will respond to changes in the temperature profile because of the water vapour feedback. If the upper troposphere were to cool, the water vapour content would therefore be expected to decrease, thus contributing to decrease the greenhouse effect. The water vapour content may decrease or increase in the lower troposphere depending on the balance between the competing effects of the CO₂ physiological forcing (less evaporation) and the water vapour feedback (less condensation). If the overall effect of changes in water vapour is to decrease the greenhouse effect, the temperature profile will be shifted downwards as illustrated in Fig. 11b. An increase in greenhouse effect or a decrease in the planetary albedo will shift the temperature profile upwards (Fig. 11c). The water vapour content can then increase throughout the troposphere over land in response to the CO₂ physiological

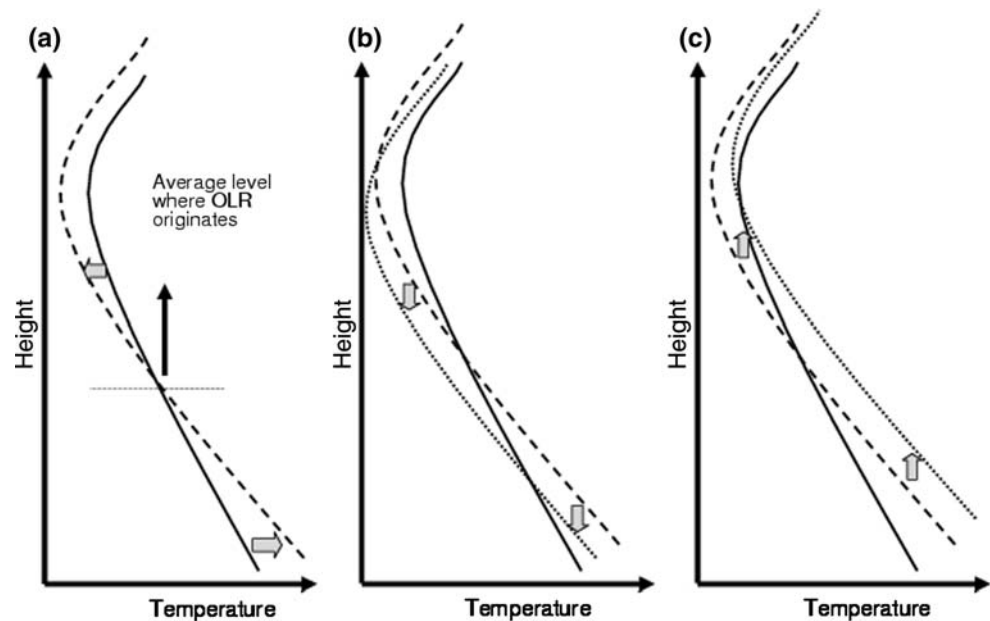
forcing. The tropospheric increase in water vapour content shows that the water vapour feedback more than compensates for the suppression in the ET flux, thus increasing the greenhouse effect. In summary, both the observed reduction in low-level cloudiness and snow cover contribute to decrease the planetary albedo, thus increasing the temperature throughout the troposphere, which then triggers an increased greenhouse effect from the water vapour feedback. The climate model results are therefore consistent with the conceptual model in Fig. 11c. A breakdown of the planetary albedo into its clear-sky and cloudy-sky components suggests that changes in cloudiness explain more than 90% of the change in planetary albedo due to the CO₂ physiological forcing, with changes in snow cover explaining less than 10%.

We now discuss how the conceptual model described above may explain the response to the irrigation forcing in BMM04. The lapse rate was found to decrease with increasing irrigation. However, there are differences with respect to the neutral point. Whereas the RADPHYS experiment warms throughout the troposphere in response to the CO₂ physiological forcing, Fig. 3b in BMM04 clearly shows a neutral point at 700 hPa when the change in tropospheric temperature is averaged over land. It is noteworthy that the irrigation forcing is mostly located in the subtropics, which may explain some of the difference with the experiments described here. In particular, we do not expect the snow albedo feedback to play a role in the irrigation experiments, thus limiting the change in planetary albedo to changes in cloudiness. The water vapour response is also different in the irrigation experiment of BMM04 (see their Fig. 2a), where the temperature decrease is associated with an increase in the water vapour content. The irrigation forcing is located primarily in dry regions, where the atmosphere is far from saturation, whereas the CO₂ physiological forcing is expected to be less important in these regions because the ET flux may already be limited by the availability of soil water. Therefore the water vapour feedback is expected to play much less of a role in the irrigation experiments than in the RADPHYS experiment. In simple terms, the change in atmospheric water vapour reservoir is more controlled by the sink in the RADPHYS experiment and more controlled by the source in the irrigation experiment.

4.5 Physiological forcing and climate sensitivity

The land-sea contrast simulated by HadCM3 is in the high range of land-sea surface warming ratios reported by Joshi et al. (2007) using the same definition for a set of IPCC models. As already discussed in Joshi et al. (2007), the large land-sea warming ratio in the HadCM3 model can be partly attributed to the CO₂ physiological forcing.

Fig. 11 Schematic representation of the change in the unperturbed vertical temperature profile (*solid line*). **a** The reduced latent heat flux at the surface leads to an increase the lapse rate (*dashed line*) around an average level assuming constant greenhouse effect and planetary albedo. **b** A decrease in greenhouse effect contributes to shift the temperature profile downwards (*dotted line*). **c** An increase in greenhouse effect and a decrease in the planetary albedo will both contribute to shift the temperature profile upwards (*dotted line*)



The additional surface warming associated with the CO₂ physiological forcing raises an interesting dilemma on how to define climate sensitivity and climate efficacy. Strictly speaking the climate sensitivity is defined as the surface temperature increase at equilibrium for a unit radiative forcing and has unit of $\text{KW}^{-1} \text{m}^2$. However, in practice the climate sensitivity is often estimated in a $2 \times \text{CO}_2$ equilibrium experiment. The climate sensitivity is then expressed in units of K for a doubling of the CO₂ concentration or it is normalised by the CO₂ radiative forcing (roughly 3.7 W m^{-2}) and expressed in $\text{KW}^{-1} \text{m}^2$. In this case there is an ambiguity as to whether the climate sensitivity does or does not include the warming effect associated with the CO₂ physiological forcing. Such an effect would have to be switched off in the $2 \times \text{CO}_2$ experiment if the climate sensitivity is to be strictly defined against a radiative forcing. Climate efficacy of a particular forcing agent is defined as “the global mean temperature change per unit forcing produced by the forcing agent relative to the response produced by a standard CO₂ forcing from the same initial climate state” (Hansen et al. 2005). The climate efficacy of other radiative forcings will therefore be lower if the CO₂ forcing agent is allowed to impact plant stomatal aperture as well as the radiative balance. Our calculations indicate that the effect is of the order of 15% in the HadCM3LC model.

5 Conclusions

We have analysed two transient climate simulations for the twenty-first century to isolate the climate response to

the CO₂ physiological forcing. The climate response to the CO₂ physiological forcing appears to be quite different to the response to the CO₂ radiative forcing. There are some indications that some of the feedbacks at work behave differently over land because of the difference in water availability. However this would need to be confirmed by a detailed feedback analysis. The land surface warming associated with the decreased ET rate is accompanied by an increase in the atmospheric lapse rate and an increase in specific humidity. The change in precipitation proved not to be statistically significant. The snow-albedo response is stronger in response to the CO₂ physiological forcing than in response to the CO₂ radiative forcing. Changes in the snow albedo, water vapour and cloud feedbacks all contribute to further warm the climate system. The CO₂ physiological forcing is one of many climate forcing mechanisms which cannot be captured by the radiative forcing concept. Earth system models are a critical tool to study the climate responses to these forcings in a more integrated way. In particular changes in vegetation dynamics and their impacts on the atmosphere also have to be accounted for as they can offset some of the impacts discussed in this study.

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