Hadley Centre
for Climate Predictions and Research

The time dependence of climate sensitivity

by:

C. A. Senior and J. F. B. Mitchell

HCTN 15

Hadley Centre Technical Note

January 2000

The Met.Office

HADLEY CENTRE TECHNICAL NOTE NO. 15

THE TIME DEPENDENCE OF CLIMATE SENSITIVITY

by

Catherine A. Senior and John F. B. Mitchell

Edited version submitted to *Geophysical Research Letters*

January 2000

Hadley Centre for Climate Prediction and Research
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SY

Abstract

A doubled CO_2 coupled ocean-atmosphere experiment has been run for over 800 years. The 'effective' equilibrium climate sensitivity to a doubling of CO_2 (the equilibrium response of the model assuming the feedbacks remained constant at the value found at any given point of the transient response) is calculated throughout the run and found to display a considerable degree of time-dependence. The time-dependence is associated with differences in cloud feedback arising from inter-hemispheric temperature differences due to the slower warming rate of the Southern Ocean. The sensitivity of the coupled model is compared to the equilibrium response of a parallel mixed-layer model experiment to assess the effect of the deep ocean feedbacks on the climate response. The time-dependence of the climate response has implications for the use of simpler models in scaling GCM results to different scenarios of forcing.

1 Introduction

The concept of the climate sensitivity, T_{eq} defined as the equilibrium global-mean temperature response to a doubling of atmospheric CO_2, is used to help compare the responses from different climate models to increases in radiative forcing, and in particular those associated with human activity. As current estimates vary by a factor of three or so (e.g. Kattenberg *et al* (1996)) this provides a useful gross measure of a model's behaviour. One can define a more general climate feedback parameter in terms of global-mean temperature response / unit global-mean radiative forcing (or its inverse as used later in this paper) (Dickinson, 1986), but as Hansen *et al* (1997) have shown, the response to a given forcing depends on the distribution of the forcing. For example, they found the largest response for forcings applied near the surface or in high latitudes.

Many estimates of climate sensitivity are based on equilibrium simulations of atmospheric general circulation models coupled to a simple mixed-layer model of the ocean (for example; Manabe and Stouffer (1980), Hansen *et al* (1984), Senior and Mitchell (1993)). More recently, atmospheric models have been coupled to dynamical ocean models enabling the estimation of the time dependent response to increases in carbon dioxide and other trace gases (e.g. Gates *et al* (1992), Kattenberg *et al* (1996)). However, as these models are computationally expensive, it is often not possible to examine the response of the climate to a large number of greenhouse gas emission scenarios. Hence simpler energy balance models can be calibrated to reproduce the global-mean coupled model results and interpolate or extrapolate them to other scenarios (e.g. Kattenberg *et al* (1996), Raper *et al* (2000)) particularly in integrated assessments of climatic impacts. These models require the specification of the climate sensitivity. There are two possible problems here. First, very few coupled models have been run to full equilibrium with doubled CO_2 concentrations. Hence, one must estimate the climate sensitivity either from the atmospheric version of the ocean-atmosphere model coupled to an oceanic mixed layer, which may respond differently due to the neglect of oceanic advection (see, for example Stouffer and Manabe (1999)), or from the transient response of the coupled model itself. The assumption in this second case is that the climate sensitivity is constant in time. It is this second issue that we address in this paper. We estimate the climate sensitivity as a function of time from a long coupled model experiment with HADCM2. We find that the effective climate sensitivity varies systematically with time and we investigate the reasons for this variation. Finally we discuss the generality of our findings and the implications for the use of simpler energy balance models to evaluate the effects of emission stabilization scenarios.

2 Model and Experimental Design

a *Model description*

The climate model used is HADCM2, a version of the Hadley Centre climate model on a 2.5° latitude by 3.75° longitude grid with 19 levels in the atmosphere and 20 levels in the ocean (Johns *et al*, 1997). The atmospheric model (HADAM2), which is a low resolution version of that used for Numerical Weather Prediction, includes an interactive land surface, a penetrative convection scheme with downdraughts and a stability-dependent boundary layer. Cloud radiative properties are a function of the model-generated cloud water content. Clouds are amalgamated into three layers (high, medium and low) for the treatment of solar radiation, but the longwave fluxes are derived from the full cloud distribution. Runoff is a function of the prescribed vegetation and soil type, and of soil moisture content and convective precipitation intensity. The ocean includes an explicit mixed layer and a simple representation of ice dynamics. A general assessment of the simulation of current climate is given in Johns *et al* (1997).

b *Experimental design*

The coupled model underwent a spin-up of 510 years during which flux adjustment terms were calculated (Johns *et al*, 1997). A control integration, initialised from the end of the spin-up, has been run for around 1800 years using pre-industrial equivalent CO_2 concentrations. 130-year averages have been taken from a period after the model has spun up. A further experiment was initiated from the end of the spin-up, in which the CO_2 concentration was increased by 1% (compounded) until it had reached double the initial value, at year 70, and was held constant thereafter. The experiment was continued for a further 830 years and decadal averages from the period after the time of doubling are used in this paper. Flux adjustments were applied identically to both control and 2xCO_2 experiments. A slow drift of 0.016K century^{-1} in surface temperature in the control experiment was allowed for in the calculation of the transient response.
HADAM2 was also coupled to a 50-m mixed-layer (or 'slab') ocean. This model (HADSM2) was run to equilibrium with both pre-industrial CO_2 concentrations and with twice pre-industrial values (2xCO_2) for 30 years and averages were taken over the final twenty years in each case.

3 Climate Sensitivity Analysis

The energy balance of the climate system can be represented by a simple zero-dimensional model e.g. Dickinson (1986);

$$C\frac{\partial \Delta T}{\partial t} + \lambda \Delta T = G_0 \tag{1}$$

where C is the heat capacity of the system, G_0 is the instantaneous net downward flux at the tropopause due to a given forcing, e.g. a doubling of CO_2, ΔT is the surface temperature response and λ is a climate feedback parameter. At equilibrium λ is given by

$$\lambda = \frac{G_0}{\Delta T_{eq}} \tag{2}$$

The calculated instantaneous flux at the tropopause for a doubling of CO_2 in HADCM2 after allowing for the stratosphere to come into thermal equilibrium is 3.47 Wm^{-2}. As the stratosphere is then in equilibrium, this is then also the change at the Top of the Atmosphere (TOA) and we can redefine G_0 to be at the TOA for simplicity. Thus G_0 = 3.47 Wm^{-2}. When the concentration of CO_2 is held constant at twice the original value the global-mean temperature continues to rise (Figure 1a; solid line). The initial imbalance of 1.2 Wm^{-2} in the net radiative flux at the TOA drops to less than 0.4 Wm^{-2} over the next 830 years (Figure 1a; dashed line), giving a slow trend of about 0.1K $century^{-1}$ over the last 500 years. We can make an estimate of the equilibrium climate sensitivity by calculating the time-dependent 'effective' climate feedback parameter, λ_{eff} (related to the effective climate sensitivity (Murphy, 1995), $\Delta T_{eff} = \frac{G_0}{\lambda_{eff}}$) where;

$$\lambda_{eff} = \frac{G_0 - \Delta(Q-F)_t}{\Delta T_t} \tag{3}$$

Q and F are the net downward shortwave and net outgoing longwave fluxes at the TOA and $\Delta(Q - F)_t$ is the remaining net TOA radiative perturbation at a given time, t. The effective climate sensitivity at a given time can be thought of as 'the climate sensitivity due to a doubling of CO_2 that would occur if the AOGCM ran to equilibrium with feedback strengths held fixed at the values diagnosed at that time during the run'. λ_{eff} decreases over the length of the $2xCO_2$ experiment (Figure 1b: thick solid line) giving an increase in the effective climate sensitivity from around 2.6K when CO_2 is first stabilized, to nearly 3.8K at the end of the experiment. This implies that the feedback strengths are not constant in time and that the effective climate sensitivity calculated at the start of the integration is not a good estimate of the "equilibrium value". The true equilibrium climate sensitivity of the coupled model remains unknown, but the equilibrium temperature change, ΔT_{eq}, of a parallel mixed-layer integration with identical atmospheric physics is 4.1K

Some insight into the reason for the increase in ΔT_{eff} with time can be gained by considering λ_{eff} separately for clear and cloudy skies. We assume that the individual feedback contributions to the overall climate sensitivity are linear and independent (Hansen *et al*, 1984). We define the clear-sky effective climate sensitivity as the effective sensitivity of the whole system after removing the effect of clouds. So;

$$\lambda^c_{eff} = \frac{G_0 - \Delta(Q_c - F_c)_t}{\Delta T_t} \tag{4}$$

and the cloud contribution to the feedbacks by

$$\lambda^{cl}_{eff} = \frac{-\Delta(CF)_t}{\Delta T_t} \tag{5}$$

where CF is the cloud radiative forcing (e.g. Ramanathan *et al* (1989)) and;

$$\Delta(CF)_t = \Delta(Q - Q_c)_t - \Delta(F - F_c)_t \tag{6}$$

The clear sky value, λ^c_{eff} (Figure 1b; thick dashed line) is almost constant (there is a small positive trend which would tend to decrease ΔT_{eff} with time). Thus, the clear sky feedbacks (including those due to water vapour, lapse rate and surface albedo) change little as the model approaches equilibrium. Much of the slight increase in λ^c_{eff} found here comes from changes in the shortwave flux consistent with a slight decrease in surface albedo feedbacks, which might be expected as the model warms due to the decrease in the area of snow and ice (Spelman and Manabe, 1984). However, changes in the character of cloud feedbacks must be mainly responsible for the trend in our estimates of λ_{eff} and ΔT_{eff}.

We can compare λ_{eff} and λ^c_{eff}, with the all-sky and clear-sky equilibrium climate feedbacks found in the parallel mixed-layer experiment, λ_{slab} and λ^c_{slab}. λ^c_{eff} converges to λ^c_{slab} (Figure 1b),

suggesting this is the equilibrium clear-sky feedback in the coupled model. In contrast, λ_{eff} has not reached the equlibrium value, λ_{slab} after more than 800 years. It is possible that λ_{eff} and λ_{slab} may not be identical even at full equilibrium, as the mixed-layer model does not include feedbacks from changes in the deep ocean circulation, although other authors (Stouffer and Manabe (1999), Watterson (1999)) have found that the effective climate sensitivity from AOGCMs is typically close to the equilibrium result taken from mixed-layer experiments.
The cloud feedback can be usefully further split into the shortwave and longwave components;

$$\lambda_{eff}^{swcl} = \frac{\Delta(Q - Q_c)_t}{\Delta T} \tag{7}$$

$$\lambda_{eff}^{lwcl} = \frac{-\Delta(F - F_c)_t}{\Delta T} \tag{8}$$

respectively. Note the signs have been chosen so that a positive (negative) value implies a positive (negative) cloud feedback ie a magnified (reduced) warming). λ_{eff}^{swcl} responds quickly initially but changes little after about 300 years whilst λ_{eff}^{lwcl} changes more slowly initially but has not reached equilibrium after 830 years (Figure 1c). The time dependence of the cloud feedbacks may arise from changes in either macrophysical or microphysical cloud properties. These can be separated by considering the changes in cloud amount / height and cloud radiative properties throughout the integration.

4 Mechanisms of cloud feedback

As CO_2 increases, the near-surface temperature in the Northern Hemisphere warms relative to the Southern Hemisphere due to the inertia of the Southern ocean in the transient part of the coupled experiment. At upper levels, however, the warming is more uniform and the temperature lapse between the surface and 200Hpa is reduced in the Southern Hemisphere early in the experiment (Figure 2; dashed line).
The reduced lapse rate increases the static stability in the Southern Hemisphere and convective activity is weakened in the transient part of the experiment. As a result deep convective cloud decreases much more markedly than in the Northern Hemisphere (Figure 3a). Conversely, stratiform cloud reduces less than in the Northern Hemisphere because of an increased tendency of moisture to be trapped in the more stable boundary layer (Figure 3b). The low cloud acts to cool through reflection of solar radiation, whilst the reduced deep convective cloud has a smaller greenhouse effect. As our 2xCO_2 experiment continues to near-equilibrium, the Southern Hemisphere warming 'catches up' with the Northern Hemisphere and these assymetries reduce and even reverse. In the Northern Hemisphere, the temperature lapse rate increases slightly initially as the land and parts of the ocean surface warm quickly, but the effect of moist processes at upper levels, especially in the tropics, tends to reduce the lapse rate and so increase the stability / degree warming with increasing temperature (Figure 2; solid line). It is worth noting that if the moist processes are similarly important in the Southern Hemisphere then the trend to de-stabilisation seen in Figure 2 is even more striking.
The changes in convective and low stratiform cloud in the Northern Hemisphere (Figure 3; solid lines) and at other levels in both hemispheres (not shown) have no time dependence. The Southern hemisphere cloud changes are consistent with an increasingly positive cloud feedback as seen in the trends in λ_{eff}^{swcl} and λ_{eff}^{lwcl}.
In the Southern Hemisphere, the pattern of zonally averaged cloud changes at the time of doubling (Figure 4a) is very different to that in the mixed-layer model (Figure 4c), as was found

by Murphy and Mitchell (1995), but by the end of the experiment the mixed-layer and coupled model patterns (Figure 4b) are very similar

Changes in cloud radiative properties can also affect the radiation budget and as they depend non-linearly on cloud water path are a potential source of non-constancy in the cloud feedback. However, here there is little time dependence in either hemisphere except for a slight increase in high cloud albedo in the early part of the run (Figure 5 shows the globally-averaged in-cloud albedo and emissivity timeseries available for a slightly shorter period from the later part of the experiment).

5 Model dependence of results

There is little agreement between different models on the details of cloud feedbacks (e.g. Cess *et al* (1996)). Given our findings are linked to simulated changes in cloud, are other models likely to behave in the same way? As far as we are aware, all coupled models produce a Northern-Southern Hemisphere asymmetry in warming when forced with a gradual increase in greenhouse gases, and presumably an increase in static stability over the Southern Ocean. It is less clear if all models will have reduced convective cloud and increased low cloud under these conditions, though both an earlier version of this model (Murphy and Mitchell, 1995) and a later version (Mitchell *et al*, 1998) do so.

Murphy and Mitchell (1995) found λ_{eff} calculated after 80 years to be very similar to the climate sensitivity of a lower resolution version of the atmosphere coupled to a mixed layer model, although it varies considerably earlier in the run. In HADCM2, even after more than 800 years with constant forcing, the effective climate sensitivity of the coupled model remains less than the equilibrium climate sensitivity in the parallel mixed-layer experiment. In HADCM2 the relative increase in convective cloud through the experiment is still continuing after 830 years.

Relatively few coupled experiments have been performed where the forcing has been stabilised and the model run for several centuries. Of these, at least two (Manabe and Stouffer (1994), Watterson (1999)) have used very different convection schemes to that in HAdCM2. In contrast to the results shown here, Watterson (1999) finds that the effective climate sensitivity varies very little through the coupled experiment. Hence, there is a need to verify if our findings extend to other models and to assess the realism of the contributing mechanisms. In a 4xCO_2 integration with the latest Hadley Centre model (HADCM3; Gordon *et al* (1998)) λ_{eff} also decreases in time implying that the ΔT_{eff} to a doubling of CO_2 would increase from 3.1 to 3.4K over the 180 year experiment, again due to non-constant cloud feedbacks. This compares to a rise from 2.7 to 3.2K over the same period in the HADCM2 experiment described here. The mixed layer experiment, HADSM3 has a T_{eq} of 3.3K.

6 Concluding remarks

If the effective climate sensitivity does vary with time, then there are some important consequences for model estimates of future anthropogenic climate change. First, estimating the climate sensitivity during the period when the greenhouse gas forcing is increasing will lead to an underestimation of the equilibrium sensitivity and the long term climatic effects. Secondly, the use of a single climate sensitivity in an energy balance model (such as those used in integrated assessment studies) to evaluate the effect of stabilization of emissions will lead to error. If the energy balance model uses a climate sensitivity estimated from the equilibrium response, then it will exaggerate the rate of warming when the radiative forcing is increasing. If the climate sensitivity is chosen to produce the correct rate of warming when the radiative forcing is increasing, then the long term equilibrium warming will be underestimated. A further complication is the influence of factors other than well-mixed greenhouse gases. In particular, the scattering of sunlight due to increases in sulphate aerosols which has occurred principally in the Northern Hemisphere will tend to reduce the inter-hemispheric asymmetry producing some of the change in ΔT_{eff} noted here. However, it is likely that sulphur emissions will stabilize more rapidly than greenhouse gas emissions, so this effect may become less important with time.

7 Acknowledgements

This work was funded under the U.K. Public Meteorological Service Research and Development Programme. Thanks are due to Jonathan Gregory and James Murphy for helpful discussions and comments on this manuscript.

References

Cess, R. D., M. H. Zhang, W. J. Ingram, G. L. Potter, V. Alekseev, H. W. Barker, E. Cohen-Solal, R. A. Colman, D. A. Dazlich, A. D. Del Genio, M. R. Dix, V. Dymnikov, M. Esch, L. D. Fowler, J. R. Fraser, V. Galin, W. L. Gates, J. J. Hack, J. T. Kiehl, H. L. Treut, K. K. W. Lo, B. J. McAvaney, V. P. Meleshko, J. J. Morcrette, D. A. Randall, E. Roeckner, J. F. Royer, M. E. Schelsinger, P. V. Sporyshev, B. Timbal, E. M. Volodin, K. E. Taylor, W. Wang, and R. T. Wetherald, 1996: Cloud feedback in atmospheric general circulation m odels: An update. *J. Geophys. Res.*, **101**, 12791–12794.

Dickinson, R. E., 1986: How will climate change? *The Greenhouse Effect, Climate Change and Ecosystems*, Bolin, B., B. R. Doos, J. Jaeger, and R. A. Warrick, Eds., SCOPE Report 29, John Wiley and Sons, 206–270.

Gates, W. L., J. F. B. Mitchell, G. J. Boer, U. Cubasch, and V. P. Meleshko, 1992: Climate modelling, climate prediction and model validation. *Climate change 1992: the supplementary report to the IPCC scientific assessment*, Houghton, J. T., B. A. Callander, and S. K. Varney, Eds., Cambridge University Press, 97–134.

Gordon, C., C. Cooper, C. A. Senior, H. Banks, J. M. Gregory, T. C. Johns, J. F. B. Mitchell, and R. A. Wood, 1998: The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments. Hadley Centre Technical Note 1, Hadley Centre for Climate Prediction and Research.

Hansen, J. E., A. Lacis, D. Rind, L. Russell, et al., 1984: Climate sensitivity analysis of feedback mechanisms. *Climate Processes and Climate Sensitivity*, Hansen, J. E., and T. Takahashi, Eds., Geophysical Monograph 29, American Geophysical Union, 130–163.

Hansen, J., M. Sato, and R. Reudy, 1997: Radiative forcing and climate response. *J. Geophys. Res.*, **102**(D6), 6831–6864.

Johns, T. C., R. E. Carnell, J. F. Crossley, J. M. Gregory, J. F. B. Mitchell, C. A. Senior, S. F. B. Tett, and R. A. Wood, 1997: The second Hadley Centre coupled ocean-atmosphere GCM: Model description, spinup and validation. *Climate Dynamics*, **13**, 103–134.

Kattenberg, A., F. Giorgi, H. Grassl, G. A. Meehl, J. F. B. Mitchell, R. J. Stouffer, T. Tokioka, A. J. Weaver, and T. M. L. Wigley, 1996: Climate models—projections of future climate. *Climate Change 1995. The Science of Climate Change*, Houghton, J. T., L. G. Meira Filho, B. A. Callander, N. Harris, A. Kattenberg, and K. Maskell, Eds., Cambridge University Press, 285–358.

Manabe, S., and R. J. Stouffer, 1980: Sensitivity of a global climate model to an increase of CO_2 concentration in the atmosphere. *Journal of Geophysical Research*, **85**, 5529–5554.

——, and ——, 1994: Multiple century response of a coupled ocean-atmosphere model to an increase of atmospheric carbon dioxide. *J. Climate*, **7**, 5–23.

Mitchell, J. F. B., T. C. Johns, and C. A. Senior, 1998: Transient response to increased greenhouse gases using models with and without flux adjustment. Technical report, Hadley Centre, UK Met Office, London Road, Bracknell, UK, RG12 2SZ.

Murphy, J. M., and J. F. B. Mitchell, 1995: Transient response of the Hadley Centre coupled ocean-atmosphere model to increasing carbon dioxide. Part II: Spatial and temporal structure of response. *J. Clim.*, **8**, 57–80.

——, 1995: Transient response of the Hadley Centre coupled ocean-atmosphere model to increasing carbon dioxide: Part III. Analysis of global-mean response using simple models. *J. Climate*, **8**, 496–514.

Ramanathan, V., R. D. Cess, E. F. Harrison, P. Minnis, B. R. Barkstrom, E. Ahmad, and D. Hartmann, 1989: Cloud radiative forcing and climate: Results from the Earth Radiation Budget Experiment. *Science*, **243**, 57–63.

Raper, S. C. B., J. M. Gregory, and T. J. Osborn, 2000: Use of an upwelling-diffusion energy balance climate model to simulate and diagnose A/OGCM results. *Climate Dynamics*. Submitted.

Senior, C. A., and J. F. B. Mitchell, 1993: Carbon dioxide and climate: The impact of cloud parameterization. *J. Climate*, **6**, 393–418.

Spelman, M. J., and S. Manabe, 1984: Influence of oceanic heat transport upon the sensitivity of a model climate. *Journal of Geophysical Research*, **89**(C1), 571–586.

Stouffer, R. J., and S. Manabe, 1999: Response of a coupled ocean-atmosphere model to increasing atmospheric carbon dioxide: Sensitivity to the rate of increase. *J. Climate in Press*.

Watterson, I. G., 1999: Interpretation of simulated global warming using a simple model. *Accepted for publication in J. Climate*.

Wetherald, R. T., and S. Manabe, 1988: Cloud feedback processes in a general circulation model. *J. Atmos. Sci.*, **45**, 1397–1415.

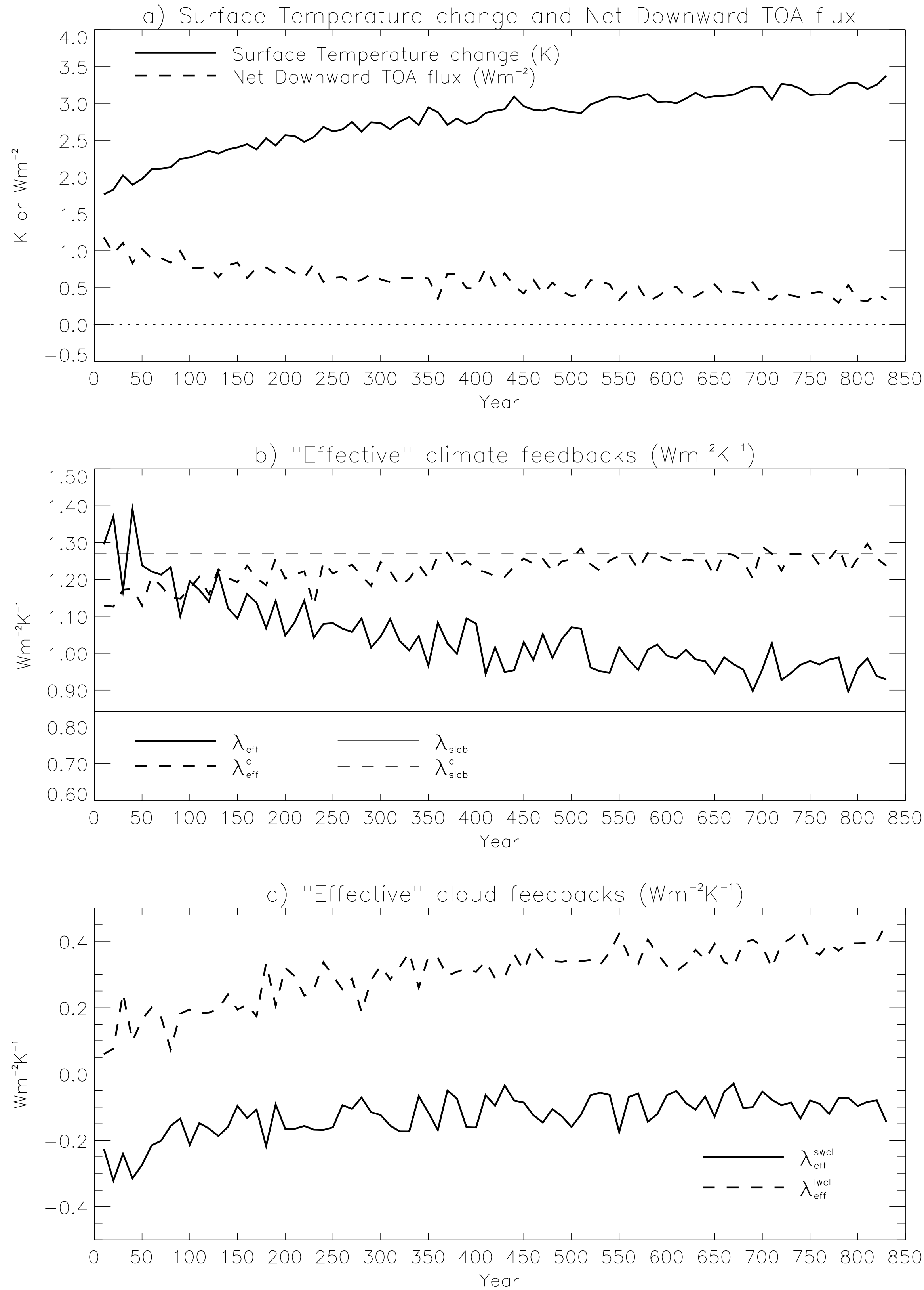

Figure 1: Decadal averages of a) temperature response K (solid line) and net downward heat flux at the top of the atmosphere Wm^{-2} (dashed line); b) λ_{eff}, (thick solid line), λ^{c}_{eff} (thick dashed line) and the equilibrium values, λ_{slab} (thin solid line), λ^{c}_{slab} (thin dashed line). All in Wm^{-2}K^{-1}; c) λ^{swcl}_{eff} (solid line) and λ^{lwcl}_{eff} (dashed line), both in Wm^{-2}K^{-1}

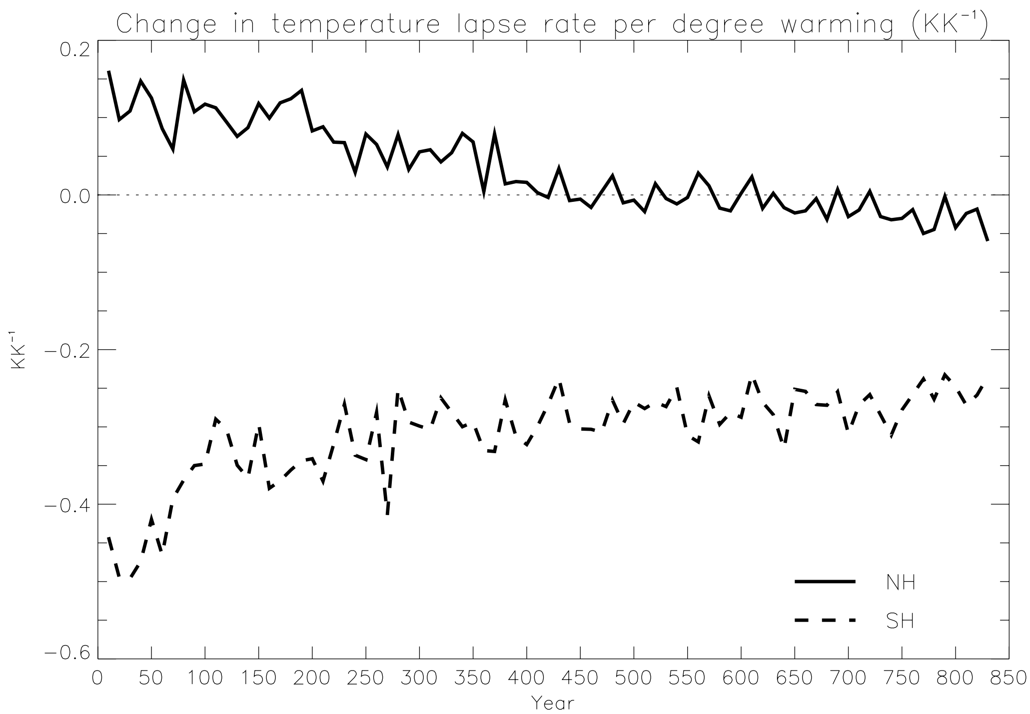

Figure 2: Decadal changes in temperature lapse rate (surface-200Hpa) / degree warming in each hemisphere (KK^{-1})

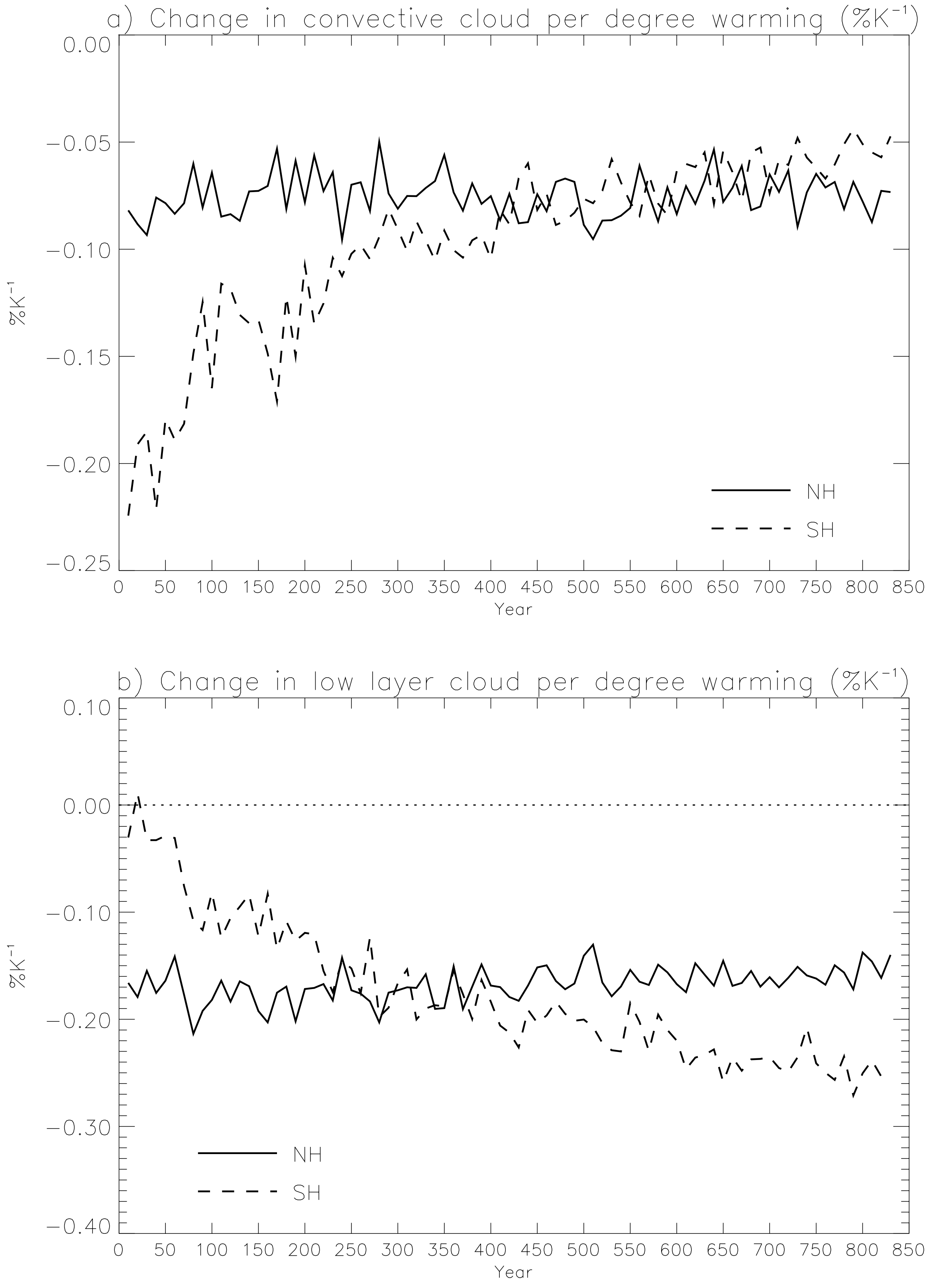

Figure 3: Decadal averages of cloud changes / degree warming (%K^{-1}) in the Northern (solid line) and Southern (dashed line) Hemispheres for; a) Convective cloud; b) Low layer cloud (levels 1-4)

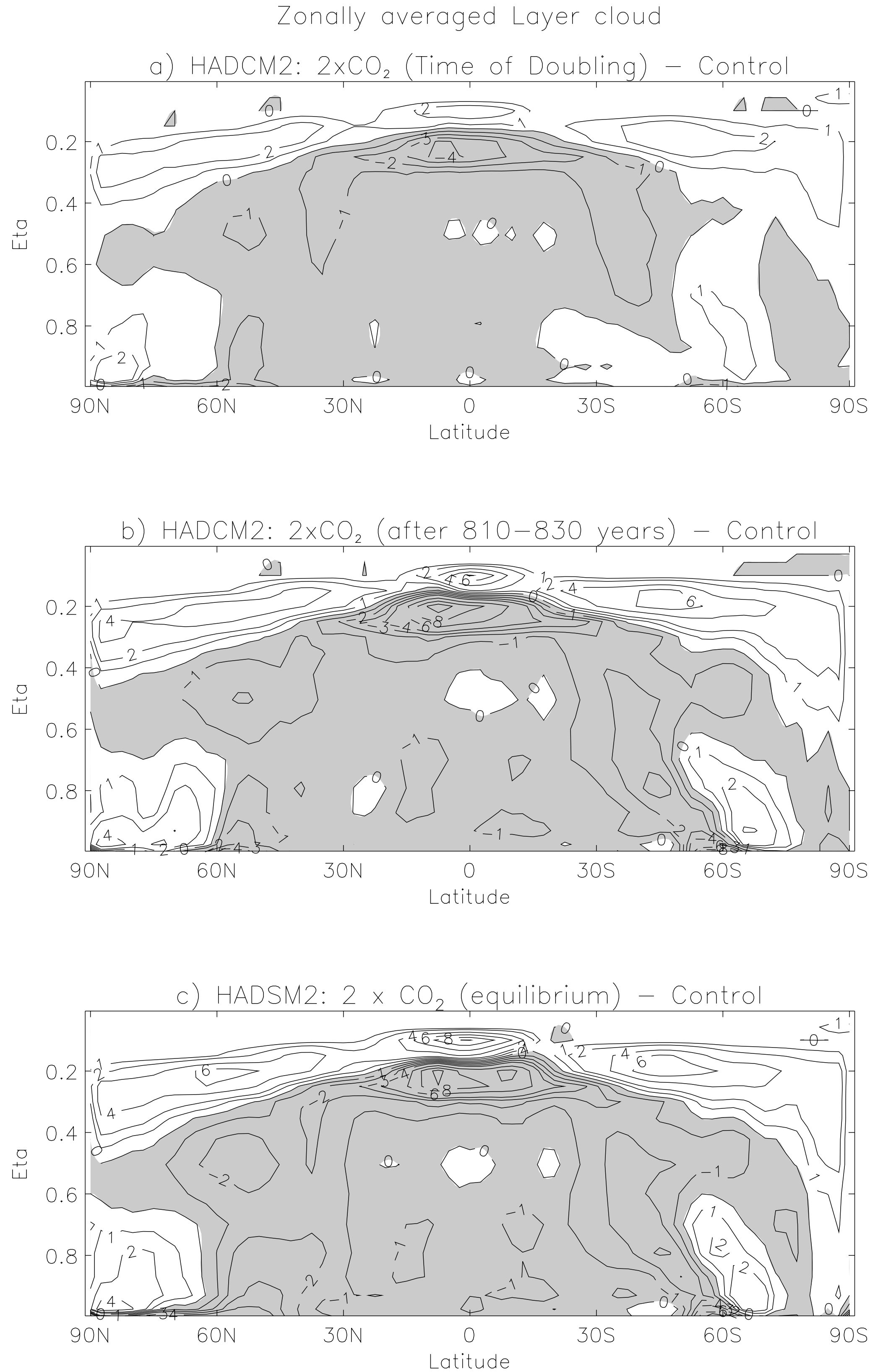

Figure 4: Zonally averaged cross-sectional changes in layer cloud relative to the control as a function of height due to a doubling of CO_2. Contours are at 0,$\pm$ 1,2,3,4 and every 2%. a) HADCM2: 20 year mean around the time of doubling (year 70) - control; b) as (a) but for a 20 year mean after 810-830 years - control; c) HADSM2: 20 year mean at equilibrium - control.

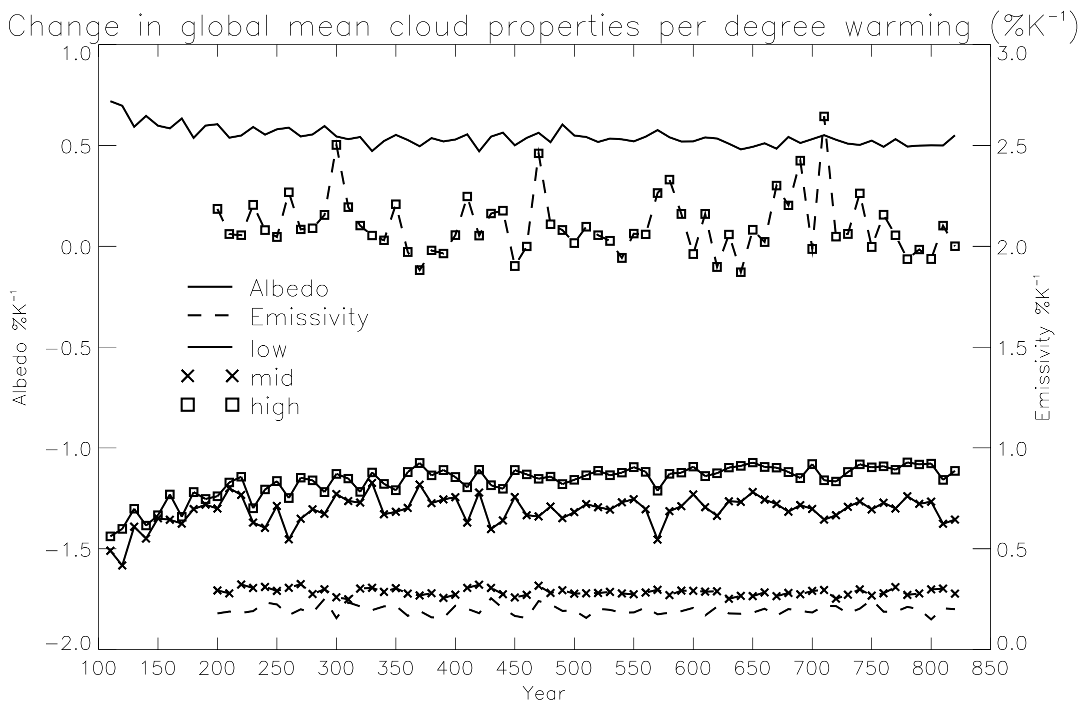

Figure 5: Decadal averages of changes in global-mean cloud radiative properties / degree warming ($\%K^{-1}$). Solid lines: In-cloud albedo (left hand axis), Dashed Line: In-cloud emissivity (right hand axis). No symbols: low cloud, crosses: mid-cloud, boxes: high cloud.