The mechanisms of cloudiness evolution responsible for equilibrium climate sensitivity in climate model INM-CM4-8

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Key Points:

- In a climate model equilibrium climate sensitivity can be changed from 1.8 K to 4.1 K by changing cloud paramerization only.
- Processes of cloud generation and dissipation responsible for equilibrium climate sensitivity change are studied
- The increase of equilibrium climate sensitivity did not lead to increase of natural variability.

Abstract

The effect of changes in cloudiness parameterization on the equilibrium climate sensitivity (ECS) of the climate model INM-CM4-8 is investigated. The reasonable changes in parameterization of cloudiness amount only lead to variability of ECS in INM-CM4-8 in the interval of 1.8 - 4.1 K, that is more than a half of the interval for CMIP6 models. Two mechanisms are mainly responsible to increase of ECS. The first one is increase of cloudiness dissipation in warmer climate because of increased water vapor deficit in non-cloud fraction of a cell. The second one is decrease atmospheric boundary layer cloudiness generation in warmer climate. The amplitude of natural climate variability change with respect to ECS change was studied. Increase of ECS doesn't lead to increase of the value of natural climate variability at most time and spatial scales.

Plain Language Summary

One of the most important problems in present day climatology is different equilibrium climate sensitivity (ECS) for present day climate models. The diapason of uncertainty is 1.8-5.6 K. Climate model INM-CM4-8 has the lowest ECS of 1.8 K. In this study, it is shown that different assumptions about cloudiness formation only lead to change of ECS in the interval of 1.8-4.1 K that is more than a half of the interval for all present day climate models. Treatment of two mechanisms: cloud dissipation because of mixing clouds and unsaturated environment, and cloud generation in atmospheric boundary layer are mainly responsible for different ECS. Present day climate models with high ECS usually show also higher natural climate variability than models with low ECS. But in INM-CM4-8 increase of ECS doesn't lead to increase of natural variability in global mean annual temperature and other climate indicators.

1 Introduction

Equilibrium climate sensitivity is one of the most important parameters of a climate model. Usually it is defined as equilibrium global mean near surface temperature change after doubling of CO_2 concentration. The interval of ECS for CMIP3 climate models is 2.1-4.4 K (Meehl el al. 2007). Different ECS for different climate models and impossibility to calculate in exact way ECS for real climate system became one of the most important problem of present day climatology. For CMIP5 climate models the diapason of uncertainty for ECS became approximately the same: 2.1-4.7 K (Flato et al. 2013). Further development of climate models lead to increase of uncertainty in ECS. In CMIP6 the lowest ECS is 1.8 K while the highest ECS is 5.6 K (Zelinka et al. 2020). Analysis of climate feedbacks show that main reason of uncertainty is the difference in feedback between warming and cloudiness (Zelinka et al. 2020). In the models with high ECS clouds reduce significantly during global warming, and this results in positive feedback, while the models with low sensitivity show small cloudiness changes during global warming, or even some increase of clouds in warmer climate, what means negative feedback (Bony et al. 2006). The change of low clouds in warm regions that produce strong negative cloud radiation forcing is especially important for ECS (Bony et al. 2015).

Climate models developed in the Institute of Numerical Mathematics Russian Academy of Science (INM RAS) usually had ECS at the low limit of the interval. In CMIP3, ECS of climate model INMCM3 vas equal 2.1 K; in CMIP5 ECS of climate model INMCM4 was equal 2.1 K, and in the last phase of comparison CMIP6, ECS of climate models INM-CM4-8 and INM-CM5-0 was equal approximately 1.8 K. The model CAMS-CSM1-0 has the second lowest ECS about 2.3 K. CMIP6 models differ from each other in all blocks. So it is difficult to answer such questions as: why some model has low ECS, but other model has high ECS? Or: how can we increase or reduce the ECS in some model? The experience of communication of the author with model developers show that often it is difficult to change on purpose ECS of some climate model if even, for example, such change is expected to reduce systematic model bias in reproduction of observed climate change in historical period.

Another question closely connected with the problem of climate model and real climate system ECS is the amplitude of natural climate variability. It was shown for CMIP5 model ensemble (Nijsse et al. 2019) that models with high ECS usually have higher interannual, or decadal, natural variability of global mean near surface temperature (GMST).

In this paper, we study the problem how the changes in cloudiness parameterization in climate model INM-CM4-8 influence on ECS of the model. In particular, how much we can increase ECS by changing only the cloud parameterization within reasonable limits. All other model blocks and parameterizations will be unchanged. We will check also whether increase of ECS leads to increase of natural variability in global mean and regional GMST.

We leave behind the scene such important questions as: what is correct value for ECS; which version model gives cloudiness changes in warmer climate more consistently with available observations; must climate model include listed mechanisms of cloud change in warmer climate or not; what about simulation of historical climate changes using model versions with low and high sensitivity, and so on. Each of such questions can be a subject of special study, or several studies.

2 Model and numerical experiments

Climate model INM-CM4-8 (Volodin et al. 2018) is used for numerical experiments. Standard model version used in CMIP6 includes parameterization of cloud fraction C according diagnostic relations Smagorinsky type (Galin 1998):

$$C = a\,r + b \qquad (1)$$

where r - relative humidity, a and b - coefficients that can depend some factors such as height above sea level, vertical stability, temperature and so on. Cloud water W is calculated diagnostically as some empirical function of temperature T and pressure P (Galin 1998): $W=W(T,P)$. Let's call this version of the model version 1.

In the version 2, we replaced diagnostic formulas for calculation of C and W by prognostic equations following Tiedtke (1993). Here we give only those formulas that will be directly used in this study. Equation for C is written as follows:

$$\frac{\partial C}{\partial t} = A + G_{CV} + G_{BL} + G_{H} - D + F \qquad (2)$$

Here t - time, A - term of advection, G - generation terms due to deep convection (CV), boundary layer processes (BL), large scale adiabatic and diabatic cooling or heating (H, this term can be both positive or negative); D - dissipation term due to mixing of cloud air and unsaturated environment; F - the impact of gravitational falling. Below the expressions for terms used in this study are presented.

It is assumed that generation of cloudiness by deep convection occurs at the uppermost level of convection and is proportional to upward convective mass flux at the low boundary of the level F_{CV}:

$$G_{CV} = F_{CV} / (\rho \Delta z) \quad (3)$$

where ρ is air density, and Δz is cell thickness. In other words, it is assumed that volume of uplifted air equals the increase of cloud volume in the cell. In the model, convective parameterization of Betts (1986) is used. It does not calculates convective mass flux explicitely, so, additional assumptions needed to estimate it. It is assumed that

$$F_{CV} = C_{CV} * \rho * (Z_T - Z_B) * \mathrm{Pr} / Pw \quad (4)$$

where C_{CV} - adjustable dimensionless factor of order 1, Z_T and Z_B - height of top and bottom of convection, Pr - convective precipitation intensity, Pw - precipitable water in convective column.

Similarly, it is assumed in the model that generation of cloudiness at the upper level of boundary layer is proportional to mass flux F_{BL} produced by turbulence if rising air reaches saturation:

$$G_{BL} = F_{BL} / (\rho \Delta z) \quad (5)$$

where F_{BL} is estimated from:

$$F_{BL} = C_{BL} * \rho * (Z_T - Z_B) * F_S / Pw \quad (6)$$

Here C_{BL} - adjustable dimensionless factor of order 1, Z_T and Z_B - height of top and bottom of boundary layer, F_S - surface evaporation, Pw - precipitable water in boundary layer.

At least, dissipation of clouds because of mixing with unsaturated environment is calculated as follows:

$$D = C_M (Q_{MAX}(T) - Q) / W \quad (7)$$

where C_M - adjustable factor (mixing coefficient), Q - specific humidity, $Q_{MAX}(T)$ - saturated specific humidity, W - cloud water.

Model version with prognostic calculation of cloud fraction and cloud water according to Tiedtke (1993) partially presented above is version 2. To study the dependence of ECS to different mechanisms of cloud formation and dissipation, we produced additional model versions. In the average, mass flux due to deep convection calculated using (4) should decrease in global warming conditions because precipitable water increases at the rate of about 7% K^{-1}, following the increase of saturation humidity, but precipitation usually show smaller increase: 1.5-2% K^{-1} for most climate models (Collins et al. 2013). This should produce decrease of upper tropical cloudiness in warmer climate. Assuming generally positive cloud radiation forcing of upper tropical clouds, this mechanism should lead to decrease ECS. To estimate the impact of this mechanism on ECS, we define model version 3 identical to version 2, but value of Pw in formula 4 is prescribed as 50 kg m^{-2}, typical value for tropical convection conditions. The choice of such numerical value keeps average tropical cloudiness amount in control run at approximately the same value as in version 3. One should expect that version 3 should have higher ECS than version 2.

Generation of boundary layer clouds defined in (5), (6) also should decrease in global warming conditions. The reason is similar: precipitable water in boundary layer increases at a rate about 7% K^{-1} following increase of saturation humidity, while latent heat flux growth is not so fast: 1.5-2 % K^{-1}. This mechanism should lead to increase the ECS. To estimate the impact of this mechanism to ECS change model version 4 is introduced. It is identical to version 2, but in formula (6) term $P_W/(Z_T-Z_B)$ is replaced by mean value of 0.01 kg m^{-3} to avoid the growth of this term in warmer climate. One should expect that ECS of version 4 is smaller than that of version 2.

Cloud dissipation because of mixing of cloud air and unsaturated environment calculated according to formula (7) should increase in warmer climate because specific humidity deficit in the numerator is expected to increase in assumption of small changes of relative humidity. While the changes of cloud water in denominator are expected to be smaller. Such mechanism should lead to decrease of cloudiness at all levels in warmer climate and therefore increase of ECS. To estimate the impact of this mechanism in change of ECS we introduce model version 5. It is identical to version 4, but in (7) expression QMAX(T)-Q has been replaced by a constant value of 0.001. This numerical value was chosen so that averaged total cloudiness in version 5 is approximately the same as in version 4. It should be expected that the ECS of version 5 will be less than in version 4, and even more so in version 2.

Short information about model versions participated in the experiments for calculation ECS is summarized in Table 1.

The method of ECS estimation is commonly used in CMIP5 and CMIP6 and proposed in Gregory et al. (2004). Two model runs are preformed with a model version: control run, where all forcings are fixed at preindustrial level, and run where concentration of CO_2 in the atmosphere is 4 times higher than in control run ($4CO_2$ run). Initial state for both runs is the same and is taken from long enough control run. The length of each run is 150 years. After this, global mean difference of GMST and heat balance at the top of atmosphere (THB) for $4CO_2$ and control run are calculated for each model year. Then, data for each year are plotted at so called Gregory plots, where X-axis represents GMST difference, and Y-axis represents THB. Parameters of the best fit linear least square line are calculated. The y-intercept of the line divided by 2 provides an estimate of the effective radiative forcing from CO_2 doubling (ERF), the slope of the line provides an estimate of the net climate feedback parameter (λ), and the x-intercept of the line divided by 2 provides an estimate of the (ECS). Such plots for all CMIP6 climate models can be found in supplementary materials to Zelinka et al. (2020). Data of ERF and ECS are divided by 2 assuming that the both values for doubling of CO_2 are two times smaller than that for quadrupling of CO_2.

3 Results

The results of sensitivity experiments performed with five climate model versions are summarized in Table 1. Version 1 shows a very low ECS of 1.8 K. The reasons are both low value of negative climate feedback parameter equal minus 1.46 W m^{-2} K^{-1} (diapason minus 0.6 - minus 1.8 W m^{-2} K^{-1} for CMIP5 and CMIP6), and low value of ERF equal 2.7 W m^{-2} (diapason 2.6-4.4 W m^{-2} for CMIP5 and 2.7-4.3 W m^{-2} for CMIP6 (Zelinka et al. 2020)). Low ECS accompanied by increase total cloudiness and in particular low cloudiness in global warming conditions. The changes of both shortwave and longwave cloud radiation forcing (CRF) are negative.

Replacement of parameterization scheme for cloudiness in version 2 changes all parameters dramatically. ECS more than doubles to 3.8 K. ERF increases to 3.8 W m^{-2} in spite there are no changes in radiation code. Climate feedback parameter increases to -1.0 W m^{-2} K^{-1}. In version 2, global warming is associated with decrease of cloudiness at all levels. The change of total CRF and shortwave CRF after global warming became positive. The analysis of the results of sensitivity experiments with version 3-5 will help us to understand the mechanisms of such significant changes.

Version 3, where the mechanism of high tropical cloudiness decrease due to decrease of convective mass flux, is suppressed, shows higher ECS with respect to version 2: 4.1 K, but the change is not very strong. The response of high cloudiness to global warming is really decreased: -0.24% K^{-1} in version 2, and -0.05 % K^{-1} in version 3. This confirm our hypothesis that suppressing of high tropical cloudiness decrease should increase ECS, but the impact of this mechanism to ECS is noticeable but not very strong.

In version 4 ECS equals 2.9 K. It is a noticeably decrease compared with version 2. This means that the mechanism of boundary layer cloudiness decrease due to decrease of cloudiness generation by boundary layer turbulence is a crucial one for ECS. The value of low cloudiness decrease changed from -0.93 % K^{-1} in version 2 to -0.44 % K^{-1} in version 4, i.e. more than two times. But, nevertheless, we still have a noticeable decrease of low clouds as well as decrease of medium and high clouds in global warming conditions.

In version 5, where the mechanism of increased cloud dissipation in global warming conditions is suppressed, ECS is reduced to 2.5 K, and climate feedback parameter decreased to -1.56 W m^{-2} K^{-1}. In this version, there is no decrease in total cloudiness during global warming, but small increase occurs, as in version 1. The changes in low cloudiness that is the most crucial for ECS, is also near zero. Some decrease of middle cloudiness and increase of high cloudiness can be treated as upward migration of cloudiness in warmer climate. We have also negative changes of CRF in version 5, while CRF changes are positive in versions 2, 3, 4. The closeness to zero of cloudiness changes in version 5 means that all main mechanisms that are responsible for decrease of clouds in warmer climate in version 2 are suppressed.

Figure 1 illustrates cloudiness change in 4CO_2 run with respect to control run in versions 1-5 normalized by global mean near surface temperature change. Version 1 shows increase of low cloudiness and decrease of middle and high cloudiness, such cloud change is mainly responsible for low ECS. In version 2, cloudiness decreases at all levels below 300 hPa, and decrease is more pronounced than that in version 1. The difference between version 3 and 2 can be seen above 600 hPa: cloudiness change in version 3 is less negative or more positive. Below 600 hPa the results of version 2 and 3 are almost identical. In version 4, we have less decrease of clouds mainly at low levels (700-1000 hPa) that leads to decrease of ECS. In version 5, decrease of cloudiness at all levels below 300 hPa is suppressed with respect to version 4. This leads to decrease of ECS.

Another point of this study is comparison of the amplitude of natural climate variability in the versions with low and high sensitivity. Control runs with versions 1 and 2 have been extended to 500 years. Root mean square deviation (RMSD) of annual mean, 5-year mean and 10-year mean GMST for both model versions is presented in Table 3. One can see that the RMSD of natural variability in GMST in version 1 is exceeds that for version 2 at all considered time scales, in spite of low ECS of version 1 and high ECS of version 2. The analysis of spatial distribution of RMSD of surface temperature shows that in general the amplitude of natural variability in the version 1 is not far from that in version 2 at monthly, annual, 5 year and 10 year time scales. There are some exceptions from this rule: for example, natural variability in tropical Pacific, including El-Nino and Pacific Decadal Oscillation is higher in version 2 with respect to version 1.

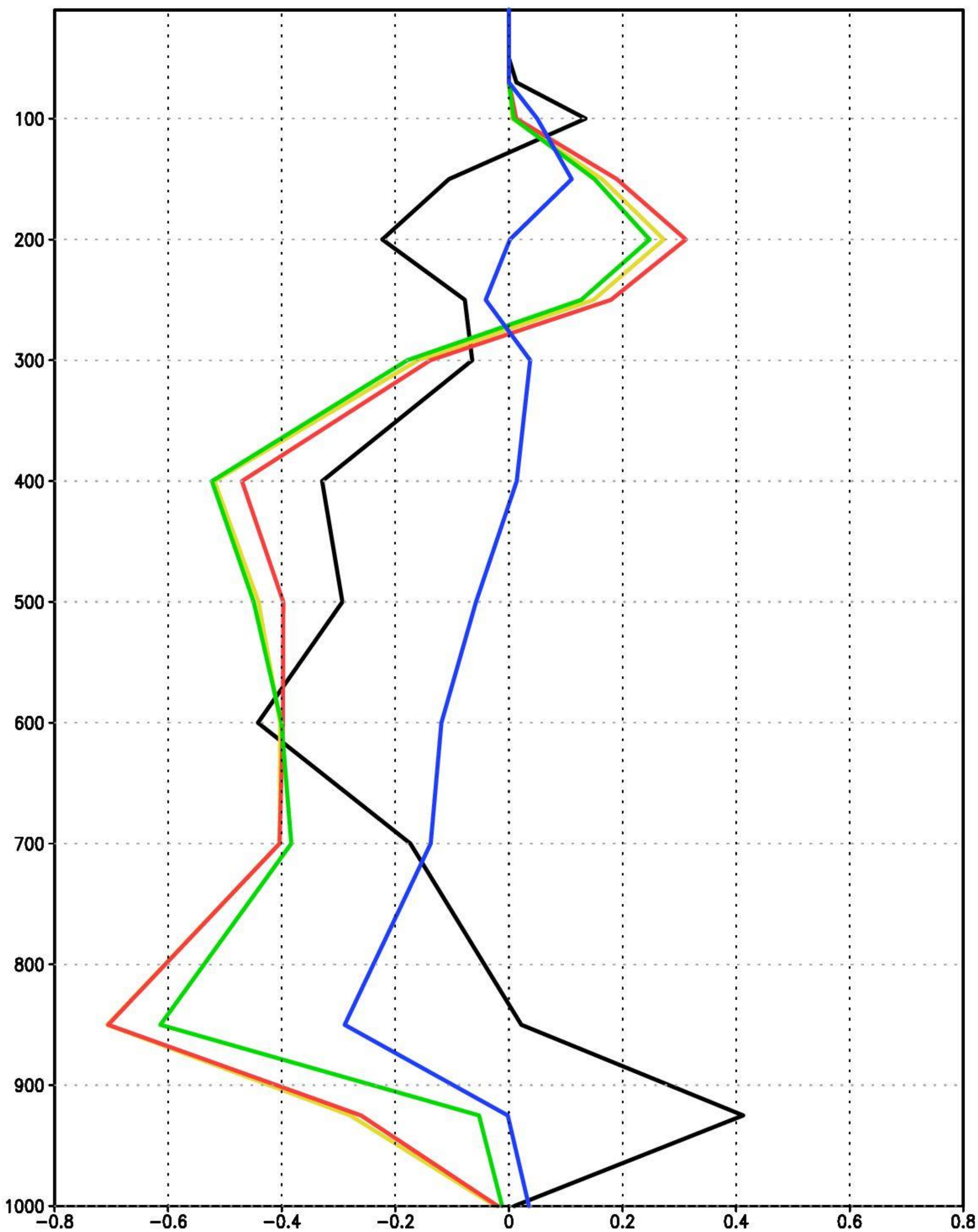

Figure 1. Cloudiness change (% K^{-1}) in $4CO_2$ run with respect to control run. Black - version 1, yellow - version 2, red - version 3, green - version 4, blue - version 5.

4 Discussion and conclusions

Climate model INM-CM4-8 has the lowest ECS equal 1.8 K among CMIP6 models. It was shown that ECS in this model can be more than doubled to 3.8 K by replacement of diagnostic cloud parameterization of Smagorinsky type by prognostic parameterization by Tiedtke (1993). In model version with low ECS we have increase of low clouds in warmer climate, while middle and upper cloudiness is slightly decreased. In the version with high sensitivity there is a pronounced decrease of cloudiness below 300 hPa in warmer climate. The mechanisms responsible for cloud decrease and their impacts to ECS are studied. Decrease of convective mass flux in warmer climate contributes to the reduction of high cloudiness in tropics. Disabling this mechanism suppress upper cloud reduction and leads to increase of ECS from 3.8 to 4.1 K. Important mechanism responsible for decrease of low clouds is the decrease of cloud formation in boundary layer by turbulent flux. Disabling this mechanism suppress low cloud reduction and leads to decrease of ECS from 3.8 to 2.9 K. Another important mechanism responsible for decrease of clouds at all levels is increased dissipation because of mixing cloud air and unsaturated environment. Increased specific humidity deficit in environment and near constant cloud water leads to increase in cloud dissipation in warmer climate at a constant mixing rate. Disabling this mechanism suppress cloud reduction at all levels and reduce ECS from 2.9 K to 2.5 K. In this version, where all listed mechanisms were disabled, there is no decrease of total cloudiness in warmer climate at all.

It should be noted that ECS in model version 1 and version 5 differ mainly because of the difference in ERF rather than the difference in climate feedback parameter. But version 1 and version 5 utilizes the same radiation block, and radiation forcing from doubling of CO_2 calculated by definition, using repeated call of radiation block in control run, but with doubling of CO_2 concentration, show value of radiation forcing about 4.0 W m^{-2} for both version 1 and version 5. The discrepancy between ERF of 2.7 W m^{-2} in version 1 and radiation forcing calculated by definition can be explained by the fact that ERF is estimated in assumption that the response of all components of climate system to CO_2 doubling is linear with respect to GMST increase. However, in reality the response to global warming can be nonlinear. In particular, in Volodin (2014) it was shown that in version 1 there is fast response of boundary layer cloudiness to doubling of CO_2. Cloudiness under boundary layer inversion increases after CO_2 doubling increases very quickly, during the first year of integration, following intensification of inversion and relatively regardless of GMST rise. In Volodin (2014) shown that this mechanism is the reason of low ERF, and that is additional source of low ECS: the impact of this mechanism to ECS is about minus 0.5 K.

Acknowledgments, Samples, and Data

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Data of control run and 4CO2 run for version 1 can be found in CMIP6 database https://esgf-data.dkrz.de/search/cmip6-dkrz/ Global and annual mean model data for versions 2-5 used here can be found in DOI: 10.5281/zenodo.4584712

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Table 1. Summary of model versions

Version	Cloud scheme	Comment
1	Diagnostic (Smagorinsky type)	
2	Prognostic (Tiedtke 1993)	
3	Prognostic (Tiedtke 1993)	As 2, but change in convective cloud formation
4	Prognostic (Tiedtke 1993)	As 2, but change in boundary layer cloud formation
5	Prognostic (Tiedtke 1993)	As 4, but change in cloud dissipation

Table 2. Equilibrium climate sensitivity ECS (K), effective radiation forcing ERF (W m^{-2}), climate feedback parameter λ (W m^{-2} K^{-1}), change of high ΔC_H, middle ΔC_M, low ΔC_L and total ΔC cloudiness (% K^{-1}), shortwave ΔCRF_{SW} , longwave ΔCRF_{LW} and total ΔCRF cloud radiation forcing (W m^{-2} K^{-1}).

Version	ECS	ERF	λ	ΔC_H	ΔC_M	ΔC_L	ΔC	ΔCRF_{SW}	ΔCRF_{LW}	ΔCRF
1	1.8	2.7	-1.46	-0.20	-0.65	0.55	0.13	-0.39	-0.55	-0.94
2	3.8	3.8	-1.00	-0.24	-0.90	-0.93	-1.01	0.92	-0.56	0.34
3	4.1	3.9	-0.95	-0.05	-0.82	-0.93	-0.84	0.80	-0.38	0.42
4	2.9	3.7	-1.28	-0.06	-1.03	-0.44	-0.50	0.52	-0.49	0.03
5	2.5	3.9	-1.56	0.18	-0.18	-0.02	0.07	0.08	-0.28	-0.20

Table 3. RMSD of annual mean, 5-year mean and 10-year mean GMST (K) for version 1 and 2.

version	1 yr	5 yr	10 yr
1	0.089	0.074	0.068
2	0.082	0.067	0.061