



INTERGOVERNMENTAL PANEL ON CLIMATE CHANGE



IPCC Working Group I
Workshop on Climate Sensitivity



École Normale Supérieure
Paris, France
26–29 July, 2004

Workshop Report

This workshop was agreed in advance as part of the IPCC workplan, but this does not imply working group or panel endorsement or approval of the proceedings or any recommendations or conclusions contained herein.

Supporting material prepared for consideration by the Intergovernmental Panel on Climate Change. This material has not been subjected to formal IPCC review processes.



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IPCC Working Group I Workshop on Climate Sensitivity

July 26–29, 2004, École Normale Supérieure, Paris, France

WGI Co-Chairs

Susan Solomon (NOAA Aeronomy Laboratory, USA)

Qin Dahe (China Meteorological Administration, China)

Scientific Steering Committee

Co-Chairs

Sandrine Bony (Laboratoire de Météorologie Dynamique du CNRS, France)

Bryant McAvaney (Bureau of Meteorology Research Centre, Australia)

Gerald Meehl (National Center for Atmospheric Research, USA)

Committee Members

George Boer (Canadian Centre for Climate Modelling and Analysis, Canada)

Chen Dehui (China Meteorological Administration, China)

William Collins (National Center for Atmospheric Research, USA)

Ulrich Cubasch (Freie Universität Berlin, Germany)

Jonathan Gregory (University of Reading, UK)

Sylvie Joussaume (National Institute for Sciences of the Universe, France)

Jean Jouzel (Institut Pierre Simon Laplace, France)

Akio Kitoh (Meteorological Research Institute, Japan)

Hervé Le Treut (Laboratoire de Météorologie Dynamique du CNRS, France)

Luo Yong (China Meteorological Administration, China)

Valentin Meleshko (Voeikov Main Geophysical Observatory, Russia)

V. Ramaswamy (NOAA Geophysical Fluid Dynamics Laboratory, USA)

Catherine Senior (Hadley Centre for Climate Prediction and Research, Met Office, UK)

Local Organizers

Local Organizing Committee

Sandrine Bony (Laboratoire de Météorologie Dynamique du CNRS, France)

Jean Jouzel (Institut Pierre Simon Laplace, France)

Hervé Le Treut (Laboratoire de Météorologie Dynamique du CNRS, France)

Armella Longrez (Institut Pierre Simon Laplace, France)

Local Support

Sophie Barbaric (Institut Pierre Simon Laplace, France)

Laurent Bourdette (Institut Pierre Simon Laplace, France)

Virginie Metrot (Institut Pierre Simon Laplace, France)

IPCC Working Group I Technical Support Unit

Martin Manning

Tahl Kestin

Scott Longmore

Melinda Tignor

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1 Introduction

1.1 Scientific Goals and Organization of Workshop

The IPCC Working Group I (WGI) held a Workshop on Climate Sensitivity on 26–29 July 2004 in Paris, France, as a major keystone in activities preparing for the WGI contribution to the IPCC Fourth Assessment Report (AR4).

One of the most important parameters in climate science is the ‘climate sensitivity’, broadly defined as the global mean temperature change ($^{\circ}\text{C}$) for a given forcing, often that of a doubling of atmospheric carbon dioxide. Climate sensitivity has played a central role throughout the history of IPCC in interpretation of model outputs, in evaluation of future climate changes expected from various scenarios, and it is closely linked to attribution of currently observed climate changes. An ongoing challenge to models and to climate projections has been to better define this key parameter, and to understand the differences in computed values between various models. Throughout the last three IPCC assessments the climate sensitivity has been estimated as being in the range 1.5 to 4.5 $^{\circ}\text{C}$ for CO_2 doubling (i.e., uncertain by a factor of three), making this parameter central to discussions of uncertainty in climate change.

While the primary reason for the substantial range in model based estimates of climate sensitivity is widely believed to be differences in their treatment of feedbacks – particularly cloud feedbacks, systematic intercomparisons have not been done to confirm that this is so for the current generation of models. Within international climate modeling projects, the development of new models together with both formal and informal model intercomparison exercises that are currently taking place by various groups suggest that a renewed focus on the reasons for different model estimates of climate sensitivity may be particularly useful at this time.

In addition, some recent studies have suggested that new insights into the likely range of climate sensitivity may be possible through comparisons of models and observational data – both contemporary and historical or paleoclimatic. Observation/model intercomparisons were a special focus of this workshop.

Other recent studies raise issues regarding the limitations of applicability of forcing/response relationships in the climate system - such as questions regarding the degree of predictability of climate and its relevance for estimates of climate sensitivity, and the degree to which forcings such as those due to solar, well-mixed greenhouse gases, or aerosols may produce different responses. A better understanding of these questions about the interpretation of climate sensitivity would also be of benefit to the WGI AR4 and was an additional goal of the workshop.

In summary, the aims of the climate sensitivity workshop were to:

- Evaluate a range of climate model results so as to relate different climate sensitivity estimates to differences descriptions of physical processes, particularly those related to atmospheric water vapor, clouds, lapse rate changes, ocean heat uptake, treatment of evapotranspiration, land-atmosphere coupling, etc.

- Obtain a more comprehensive picture of the relationships between climate sensitivity and other model features such as resolution, numerical approach, radiative transfer parameters, etc.
- Consider how current, historical, and paleoclimatic data can aid in the determination of the likely range of climate sensitivity.
- Improve the understanding of the interpretation and limits of the climate sensitivity concept, including for example possible dependencies upon different forcing agents, predictability questions, and transient and steady-state responses.
- Start a process towards objective assessment to critically determine whether the range 1.5 to 4.5°C remains appropriate in the AR4 – e.g. by defining criteria that may assist in the evaluation of results from many different climate models.

IPCC's work on this topic has been greatly aided by a number of activities under the World Climate Research Program (WCRP) and those interactions are gratefully acknowledged. A preliminary joint expert meeting was held by WCRP/IPCC on 19–22 April 2004 in Exeter, UK, to foster explicit analysis of feedbacks, with a special focus on cloud feedbacks. The report of that expert meeting is presented in Annex 5 of this report.

Given the range of issues to be considered and the commitment that would be required from major modeling groups around the world, nominations from governments were key to the identification of worldwide scientific experts. The planning for the workshop was undertaken by a broad-based Scientific Steering Committee. The Committee invited about 100 experts (see the list of participants in Annex 4) to attend and discuss the latest advances in theory, research, and data related understanding climate sensitivity, and suggest further work that could be done to enhance the understanding of climate sensitivity. These suggestions will be very valuable as an input to the deliberations of the author teams dealing with climate sensitivity-related issues in the IPCC AR4.

The WGI Technical Support Unit provided organizational support for the meeting and produced this workshop report.

1.2 Structure of Workshop

The workshop was structured around four topics:

- Climate sensitivity from models
- Climate sensitivity from observations (including modern and paleoclimatic observations)
- Radiative transfer and forcing
- Probabilistic measures of climate sensitivity

The workshop included at least one plenary session and one breakout-group session for each of these topics, as well as an overview plenary session and a key conclusions plenary session. The workshop program can be found in Annex 1.

1.3 Guide to the Findings and Material in Report

This report is organized according to session type. In Section 2 are the summaries of the four breakout-group sessions submitted by the breakout-session chairs and rapporteurs. In the case of probabilistic measures of climate sensitivity, the summary covers both the breakout session and the Climate Sensitivity and PDFs Plenary Session. Readers interested in the major conclusions of the groups and in the associated recommendations will find that material in this section.

Sections 3 and 4 contain the rapporteurs' reports of the presentations and discussions in the plenary sessions. Extended abstracts of many of the presentations are included in Annex 2.

Section 3 covers the Introduction and Overview plenary session and the Key Conclusions plenary session. The presentations in the Introduction and Overview session provided a general introduction to climate sensitivity and related issues. In the Key Conclusions session, a group of three senior researchers provided their considered views of what they heard throughout the workshop on particular topics. Readers interested a broad summary of the workshop conclusions may find this section particularly helpful.

Finally, Section 4 covers the thematic plenary sessions that covered the workshop topics described in Section 1.2.

2 Breakout Group Reports

2.1 Climate Sensitivity from Models: Breakout Group 1

Co-chairs: Gerald Meehl and Bryant McAvaney

It was recognized that representatives from most of the global coupled modeling groups would be attending the workshop, and that this presented an opportunity to present and discuss aspects of their new models being used for the AR4. Therefore, participants in this breakout group were asked before the workshop to send or bring with them specific information regarding climate sensitivity from their new models. The questions posed to the participants were:

1. What is the equilibrium sensitivity of your model (surface air DT equilibrium $2\times\text{CO}_2$ minus control with atmosphere coupled to slab ocean)?
2. What is the transient climate response (TCR) of your model (surface air DT for years 61–80 of a transient 1% per year CO_2 increase minus control, where CO_2 doubles around year 70)?
3. What is the percent change in globally averaged precipitation in your equilibrium $2\times\text{CO}_2$ simulation compared to control with atmosphere coupled to slab ocean?
4. What is the percent change in globally averaged precipitation in your 1% CO_2 simulation (percent change for years 61–80 of a transient 1% per year CO_2 increase compared to control)?
5. How do these compare to previous versions of your model, and to the numbers given in the TAR (e.g. Fig. 9.18)?
6. Why have these changed (e.g. what factors have the greatest influence on sensitivity in your model)?

A number of modeling groups were able to supply this information or, at minimum, some information regarding their models that could related to climate sensitivity. Three figures were assembled to depict the current information from the models. Figure 1 shows a plot of the percent change in precipitation as a function of equilibrium climate response (temperature

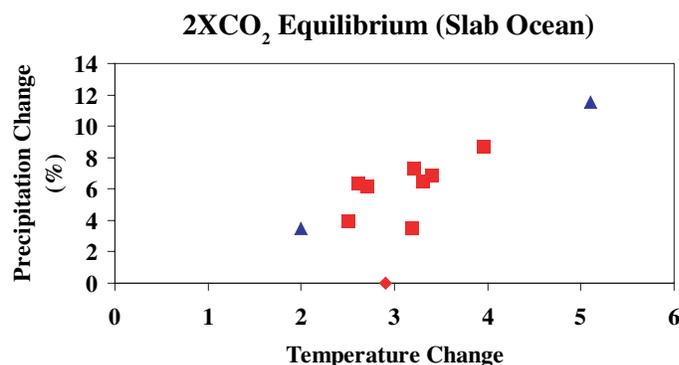


Figure 1. Change in precipitation as a function of equilibrium climate response. Blue triangles represent the range of models in the TAR (2.0–5.1°C), and red squares represent current models (2.6–4.0°C) (ECHAM4-HOPEG, “standard” CCSR-NIES, MRI, CCSM3, ECHAM5-MPIOM, GISS, GFDL CCCMA). (Red triangle is the preliminary GFDL ΔT only.)

Transient Climate Response (transient 1% CO₂ at time of doubling in a coupled model)

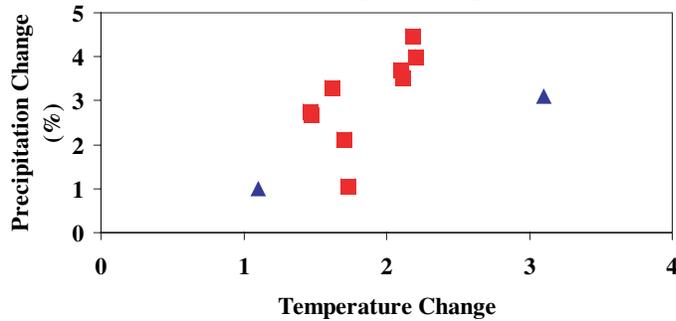


Figure 2. Change in precipitation at time of CO₂ doubling as a function of transient climate response. Blue triangles represent the range of models in the TAR (1.1–3.1°C), and red squares represent current models (1.5–2.2°C) (BCM, ECHAM4-HOPEG, “standard” CCSR-NIES, MRI, CCSM3, ECHAM5-MPIOM, IPSL, GISS HYCOM, GISS Russell).

Transient Climate Response and Equilibrium 2XCO₂ Temperature Change

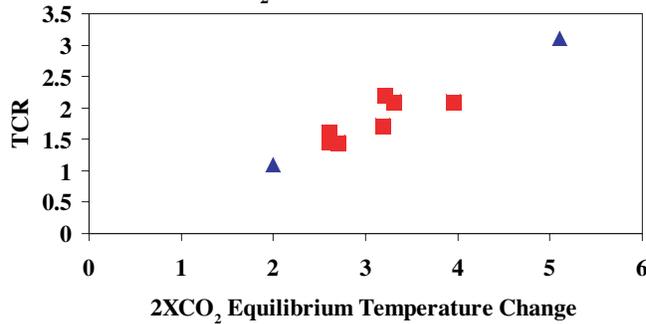


Figure 3. Equilibrium temperature change at doubled CO₂ as a function of TCR. Blue triangles represent the range of models in the TAR and red squares represent current models (ECHAM4-HOPEG, CCSR-NIES, MRI, CCSM3, ECHAM5-MPIOM, GISS HYCOM, GISS Russell).

change for an equilibrium 2×CO₂ simulation with the atmospheric model coupled to a slab ocean). Figure 2 shows a similar plot but for percent change precipitation at time of CO₂ doubling as a function of transient climate response (TCR, or the globally averaged surface temperature change at time of CO₂ doubling in a 1% per year CO₂ increase experiment with a fully coupled AOGCM). Figure 3 combines information from equilibrium doubled CO₂ as a function of TCR.

Also plotted in Figures 1 and 2 are the ranges from the models in the TAR. On first inspection, it appears that the range of responses has narrowed for the new models, and that TCR and equilibrium response may be more linearly related in this generation of models, though the breakout group agreed it is too early to make these conclusions since all models are not in hand, and higher sensitivity models cannot yet be ruled out. However, if this turns out to be the case that the range is narrowing, explicit reasons for this must be elucidated in the AR4. An initial attempt was made to discuss specific improvements to the models currently in use and represented at the workshop, and is summarized below by model:

1. *NCAR*: Improvements in prognostic cloud liquid water scheme and mid-level cloud amount compares better with observations (raised low sensitivity)
2. *GFDL*: Improved boundary layer scheme (lowered high sensitivity in latest model; many changes from TAR version difficult to attribute)

3. *MPI*: Improved and more realistic ocean model, better Southern Hemisphere simulation (raised low TCR)
4. *MRI*: Improved shortwave and longwave cloud radiative forcing (raised low sensitivity)
5. *IPSL*: many changes in all the components (atmosphere, surface, ocean, sea-ice) and great care on energy and water conservation. The changes that are supposed to have the higher impact are the new convective scheme (K. Emanuel) and the new prognostic cloud scheme (S. Bony et al.).
6. *CCCMA*: Improved cloud optical properties (raised hydrological sensitivity)
7. *CCSR-NIES*: Improved and better tuned cloud physics—cloud ice to cloud water conversion (lowered high sensitivity in standard version, different choice can produce high sensitivity in experimental version)
8. *HadGEM1*: New dynamical core, new boundary layer, new convection scheme and many other changes (slightly higher sensitivity); QUMP ensemble of slab and coupled versions of HadCM3 provide climate sensitivity probabilities as a function of parameter uncertainty
9. *CNRM*: Similar model version to TAR, no large changes in response expected
10. *BCM*: New model for AR4, improved sea ice, higher resolution ocean, no flux adjustments; note importance of initial state for climate change simulations that depends on Atlantic Ocean meridional overturning circulation oscillations
11. *MGO*: Varying atmospheric physics formulations in 9 member ensemble gives range of 2.0–4.3°C equilibrium warming
12. *ECHO-G User Group*: Using new model version ECHAM4-HOPE-G (produces mid-range sensitivities)
13. *GISS*: Model E AGCM has been coupled to two oceans, HYCOM and Russell; many model changes and difficult to attribute sensitivity change to any particular change

Improvements in various aspects of the cloud scheme were listed frequently as something that has affected climate sensitivity, though some models have made so many changes to their current versions it is difficult to ascribe any specific change to a different climate sensitivity.

General conclusions from the breakout group included:

1. It is too premature to say with any confidence that the range of sensitivity has reduced in models used for IPCC scenario simulations (all models aren't in hand yet, issues related to evaluation of large sensitivity values—see below).

2. Maximum likelihood may be a better characterization, and could be related to pdfs from perturbed physics ensembles (climateprediction.net and QUMP) or pdf estimations from the models used for IPCC scenarios (e.g. Räisänen) for probabilistic estimates, though some of the outlier estimates may not pass other metrics (see next point).
3. Need for metrics or scoring against common standard. Mean and interannual variability indices (e.g. CPI), as well as other forcing simulations (e.g. LGM, Pinatubo, or 20th century simulation) to better constrain model sensitivity
4. Is there any role of peer pressure in model sensitivity numbers? Groups don't tune for sensitivity a priori as part of model development, but they must document model improvements and performance metrics to substantiate model response (e.g. previous point)
5. Different treatments of boundary layer cloud processes appear to be important for climate sensitivity (CFMIP), and needs to be explored further.

2 Breakout Group Reports

2.2 Climate Sensitivity from Observations: Breakout Group 2

Sandrine Bony, Jonathan Gregory, Bette Otto-Bliesner and Pascale Braconnot with thanks to Alex Hall for taking notes

We need to understand better the reasons for the range of climate sensitivity displayed by current models. The main objective of this breakout group was to identify observations and diagnostics that that could lead to significant improvements in constraining and explaining this range. Discussions were organised around five time scales: last glacial maximum, Holocene, last 1200 years, instrumental period (mostly the 19–20th centuries), and satellite era (mostly the last two decades).

Studies of times before the instrumental period (together referred to as “paleoclimate”) deal with fragmentary data of only a few quantities. This hampers direct evaluation of climate sensitivity and feedbacks. Models are used to relate the available data to large-scale climate change. It is desirable for the same models to be employed as in studies of future climate change, to facilitate the use of the past to put constraints on the future. Studies of more recent periods make use of more complete data; they may enable quantitative assessment of climate sensitivity and feedbacks. Several analyses are suggested in this report. Groups are urged to contribute data from experiments recommended for AR4 and CFMIP.

Last Glacial Maximum

This period has been studied with AGCMs and AOGCMs and is a focus of PMIP 2. Global climate sensitivity can in principle be estimated from the difference between LGM and present, assuming both to be steady states, from $Q = \alpha\Delta T$, where Q is the radiative forcing, ΔT the global average surface air temperature change, and α the climate feedback parameter, related to equilibrium climate sensitivity according to $\Delta T_{2x} = Q_{2x}/\alpha$. Both Q and ΔT are subject to uncertainty, but the advantage is that the changes are large.

The important forcing is not the direct effect of insolation changes, which is minor, but the forcing resulting from the large consequential changes in various elements of the climate system, including greenhouse gases, aerosols, ice sheets and vegetation change; greenhouse gas changes are fairly well known but the others are not. An assessment of error bars on these terms would help to evaluate the implications of the uncertainty.

Uncertainty in ΔT depends quite strongly on change in tropical surface temperatures, land and ocean, because of their large area. It has been suggested that global temperature change might scale well with tropical change (Rahmstorf’s talk); however, it has only been shown within one model and not across models. PMIP 2 will compare tropical ocean and land surface temperatures estimated from paleoclimatic proxy data with those simulated by participating AOGCMs to determine the suitability of this metric for identifying outlier estimates of climate sensitivity. Other regions of scalable response will also be identified and quantified as part of the PMIP 2 model-data comparisons for the LGM.

Other aspects of LGM climate could give information about important ocean feedbacks, especially in the north Atlantic, where there could be large interrelated changes in the meridional overturning circulation, ocean temperatures and salinities, and sea ice.

There is evidence from the glacial period and during the deglaciation for abrupt climate change. If the forcings could be evaluated and global temperature estimated, perhaps these changes could be used to estimate TCR.

Holocene

PMIP 2 has a focus on the mid-Holocene (6 kyr BP), and possibly the early Holocene as well. Climate change between the mid-Holocene and the present is caused by seasonal and latitudinal changes in insolation. These generate significant responses, especially regionally, but the net radiative forcing is small. Hence the global climate change is not analogous to GHG-forced future change and global climate sensitivity cannot be evaluated (Braconnot's talk). However, important feedbacks can be studied which may contribute to climate sensitivity. Because of the nature of these feedbacks, there is regional emphasis in study of the mid-Holocene, on a range of quantities, not just mean temperature. They include changes in monsoons, interannual to multi-decadal variability, continental aridity (e.g. North America), vegetation and carbon cycle feedbacks, all of which are seen in 21st century projections as well.

Last 1200 years

Global climate sensitivity can be extracted from this period by using a climate model to predict heat uptake $N = Q - \alpha\Delta T$ given Q for a range of α and other model parameters, then comparing with the observed ΔT (and N for the recent period) to evaluate the relative likelihood of the various parameter choices (Crowley, 2000; Hegerl's talk). The advantages of this period are (a) that the relevant forcings (greenhouse gases, solar variability, volcanism, land-use change) have different time-dependence over long periods, so can be separated more effectively than for the shorter instrumental period, and (b) that aerosol forcing, which is the dominant source of uncertainty in Q for recent times, is relatively unimportant.

As for the LGM, both Q and ΔT are subject to uncertainty and would benefit from further research. Error bars should be put on the radiative forcings and the implications of the errors evaluated. Work should be done to reconcile the various global and (where possible) regional temperature reconstructions from proxy records.

Instrumental period

During the instrumental period, reasonable estimates of global temperature and some other quantities are available from direct measurement. The global climate sensitivity can be evaluated from the heat budget $N = Q - \alpha\Delta T$, but the forcings have uncertainties, which are especially serious for aerosol (TAR Fig 6.6). This has been done in a similar way to the last millennium, by comparing observations with an ensemble of model runs having a range of parameters

(Andronova and Schlesinger, 2000; Forest *et al.*, 2001) or by a straightforward calculation (Gregory *et al.*, 2002). Various analyses would improve these estimates:

- Reduced uncertainty in estimates of ocean heat uptake N from analyses of ocean temperatures (following Levitus *et al.*, 2000).
- Collation of the AOGCM radiative forcing estimates for the period in order to assess their uncertainty.
- Calculation of the uncertainty in estimated climate sensitivity that arises from the uncertainty in the forcing timeseries, especially the natural forcings and aerosols.
- Undertake AOGCM runs using separate forcings (greenhouse gases only, aerosols only, etc.) to enable statistical attribution analyses to extract the warming which can be attributed to greenhouse gases alone. The uncertainty in the estimate arising from the use of different AOGCM's "fingerprints" of the various forcing agents should be assessed. The GHG-attributable warming can then be used to estimate climate sensitivity from the global heat balance, in conjunction with the GHG forcing. The advantage of this is that the uncertainty on the GHG forcing is rather small, so a tighter constraint on ΔT_{2x} may result. On the other hand, the GHG-attributable warming is less precise than the observed total warming, since the latter has only an observational error, while the former also has the systematic uncertainty from the attribution analysis.

The following analyses do not contribute directly to evaluation of climate sensitivity but would improve models' climate change simulation:

- Ocean reanalyses and ocean observations of changes in temperature, salinity and CFCs can be used to constrain interior mixing and transport processes. This is relevant to the simulation of ocean heat uptake. All AOGCMs show less decadal variability than Levitus *et al.* (2000), and this discrepancy has to be resolved, because it may point to common deficiencies in ocean models.
- The coupled climate carbon cycle intercomparison project (C4MIP) will permit the assessment of model sensitivity of the carbon cycle to global temperature change. The carbon cycle feedback is potentially important to 21st century climate projections, but is not conventionally included in the climate sensitivity as it is not a fast feedback.

Satellite period

For this period as for the instrumental period, global climate sensitivity can perhaps be evaluated from the heat budget. The net heat flux N into the climate system is measured either by ocean heat uptake (e.g. Levitus *et al.*, 2000) or TOA net radiation from ERBE. The latter possibly has better precision for interannual variations. The estimate could be made for the period as a whole (i.e. from interannual trend and variability), the seasonal cycle (Manabe's talk) or volcanic eruptions (Robock's talk), especially Pinatubo.

In the satellite period, observations are available for a wide range of quantities with good spatial and temporal resolution. The group aimed to identify a small number of tests or metrics of processes relevant for climate sensitivity that may be applied to both observations and models (in particular within the framework of CMIP and CFMIP projects). In all cases it would be necessary to investigate whether the suggested processes do contribute significantly and in a similar way to climate sensitivity in all the models.

- Pinatubo experiments using AOGCMs or EBMs: The radiative forcing of Pinatubo eruption is the best constrained of all volcanic eruptions and the climate response was well observed. The amplitude of the climate response and the timescale of the recovery must both depend on climate sensitivity, although Wigley *et al.* (submitted paper) show that this dependence is relatively weak (because of the importance of ocean heat storage in such transient response). Pinatubo experiments may therefore set some constraint on climate sensitivity. However, they may be more useful for evaluating some feedbacks (e.g. water vapor) and their contribution to climate sensitivity than for evaluating the overall sensitivity, because other feedbacks e.g. surface albedo may not contribute to the climate response to Pinatubo in the same way as they do to GHG-forced climate change, owing to the short timescale and different geographical pattern (for example, see Figure 1). Comparing the models' response to Pinatubo would be easier if the experiment was standardised, since there is uncertainty in aerosol loading, consequent forcing, and how to translate this into a global number for use in an EBM. In the absence of standardisation, assessment of the implications of these uncertainties would be valuable.

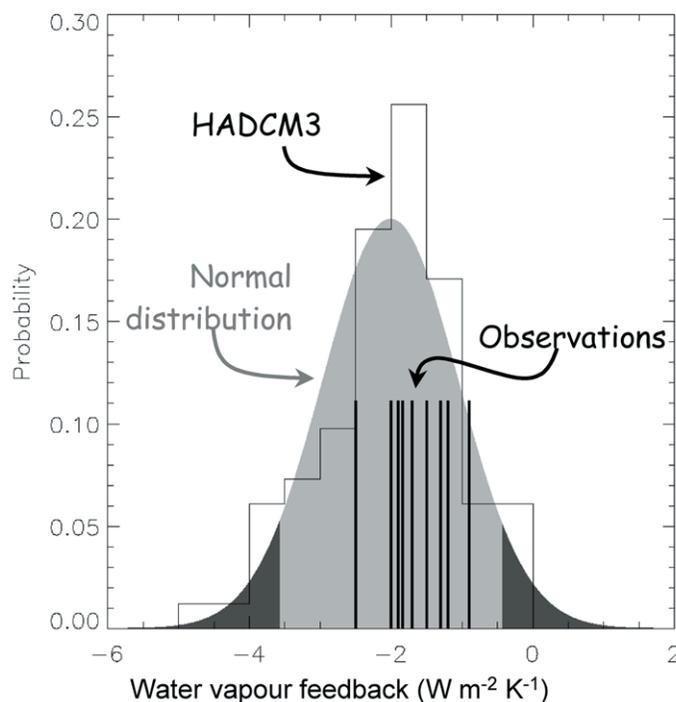


Figure 1. Estimates of the water vapour feedback from the observations and from HadCM3. The histogram is computed from 82 model estimates with a bin size of 0.5 and is shown in terms of probabilities. The shaded curve is a fitted normal distribution of model estimates with the 5% and 95% represented by darker shading. Observed estimates of water vapour feedback are indicated by the vertical lines. (From Forster and Collins, 2004)

- Seasonal and interannual variations of the radiation budget at the top of the atmosphere (in clear-sky and cloudy conditions) derived from satellites and simulated by AOGCMs or AGCMs with prescribed realistic SSTs may be analyzed to infer feedbacks involved in global climate sensitivity. As with Pinatubo forcing, the relevance of these for the long-term evolution of climate should be assessed, by comparing feedbacks in models at seasonal or interannual timescales with those for GHG-forced climate change.
- Upper tropospheric humidity (200–500 hPa) strongly affects the Earth’s greenhouse effect, particularly in subsidence regions, and interacts with upper-level cloudiness. It plays therefore a critical role in the water vapor and cloud feedbacks (Allan’s talk). The analysis of seasonal to interannual variations in upper tropospheric humidity should be pursued, using in particular the brightness temperature data at 6.7 μm derived from satellites. This would provide valuable information in particular about the variability of relative humidity in the upper troposphere, and on its potential dependence on surface temperature. Comparison with model results would help to assess the water vapor feedback and the cloud feedback associated with upper-level clouds produced by climate models.
- Boundary-layer clouds in subsidence regions have a strong impact on the Earth’s net radiation budget at the top of the atmosphere and at the ocean surface. Since large-scale subsidence affects a large area of the world, in the Tropics in particular, low-level clouds are likely to play a critical role in climate sensitivity. Relationships have been shown with model climate sensitivity. An analysis of their behaviour in observations and in coupled models, in particular their sensitivity to surface temperature and static stability changes on interannual to decadal timescales, would help to evaluate a potentially major component of cloud radiative feedback (Bony’s talk).
- Satellite observations and GCMs suggest that changes in the strength and in the frequency of baroclinic perturbations can affect the mean cloudiness, precipitation and radiation budget of midlatitude regions. An analysis of these changes in coupled models over the last decades and in global climate change simulations is encouraged. Composites of cloud and radiative properties as a function of the strength of mid-latitude storms may provide an observational constraint on cloud feedback processes in midlatitudes (Tselioudis’ talk).

In addition, some further analyses may contribute indirectly to evaluation of climate sensitivity by improving GCM performance:

- Comparing the different components of the energy budget simulated by climate models with those derived from observations both at the top of the atmosphere and at the surface would constitute a test of the models’ performance that is more constraining than just the comparison of the net radiation budget at the top of the atmosphere. ISCCP data for planetary and surface albedo, for instance, can be used to assess model variability in these quantities (Hall’s talk).
- The vertical structure of temperature changes (lapse rate, tropopause height, horizontal gradients) over recent decades can be characterized from radiosonde and satellite data. It

is suggested that these observational results be compared with coupled model results. This would be a test of the models' performance which may be a prerequisite for the credibility of the simulated lapse rate feedback.

Table 1. Summary of suggested analyses relating to quantitative evaluation of climate sensitivity or feedbacks and whether they are practicable for AR4.

Analysis	Practicable?
Undertake AOGCM runs for Last Glacial Maximum	Yes; being done with some of the AOGCMs included in AR4, although not always at same resolution.
Reduced uncertainty in heat uptake	If done by groups already working on it such as Levitus <i>et al.</i>
Collate AOGCM forcings	AOGCMs have been asked to provide time-dependent forcing for AR4 runs. It's not currently designated as a high-priority item, but without it, effective climate sensitivity cannot be calculated nor forcing uncertainty assessed. Urge modellers to provide these data (see report of Breakout Group 3).
Calculate implications of forcing uncertainty for estimate of climate sensitivity from instrumental period	Yes: ask Andronova, Forest, Gregory, Frame etc. to repeat their calculations with new estimates of forcing uncertainty.
Undertake AOGCM runs with separate forcings to calculate GHG-attributable warming	Many groups will not have enough wallclock or CPU time to provide this. Nonetheless such runs would be valuable for other attribution purposes in AR4 as well and those who may be able to do it could be encouraged to do so.
Pinatubo experiments	Being done at some level by AOGCMs for their AR4 runs if they are including volcanic forcing. Perhaps could be done by AGCMs in time.
Calculate seasonal TOA feedbacks and compare with feedbacks in climate change	Yes, from AR4 AOGCM and slab runs.
Upper tropospheric humidity variations	Yes, from AR4 AOGCM runs, using monthly and longer means of relative humidity and temperature. Simulated 6.7 μm brightness temperature is even better but probably not available.
Boundary-layer cloud in subsidence regions	Yes, could be done in CFMIP, CMIP, and AR4 runs, using seasonal and annual means of clouds and radiation.
Cloud related to mid-latitude storms	Perhaps. Daily data is required; this may be archived but transferring and processing it could take more time than is available.

2 Breakout Group Reports

2.3 Radiative Transfer and Forcing: Breakout Group 3

Co-Chairs: William Collins and V. Ramaswamy

The breakout group met to formulate recommendations and presented its findings to the final plenary on 26 July.

Accurate estimates of radiative forcing will be critical to the IPCC AR4 in order to estimate climate sensitivity and to interpret the response of climate and climate models. These estimates will be an integral part of the chapters in the AR4 concerning forcing, detection and attribution, global model evaluation, and global model projections.

Two issues are central to evaluation of radiative forcing in the AR4: first, comprehensive documentation of the forcings applied to global models used for simulations of the historical record and future scenarios; and second, evaluation of the modeled forcings for the recent past against the latest and presumably most accurate observations on global and regional scales. In turn, documentation of the forcings includes four related components:

1. common methods for computing forcing,
2. descriptions of the parameterizations for radiative transfer;
3. benchmark forcing calculations for evaluation of the radiative parameterizations in global models, and
4. information on the spatial and temporal variations in short-lived radiatively-active species.

The last objective is complicated by the fact that the modeling groups participating in AR4 will use representations of the various species that differ from model to model. The only emissions estimates supplied in the SRES scenarios are for SO₂, a chemical precursor of sulfate aerosols. Modern models often include representations of ozone, sulfate, black and organic carbon, sea salt, dust, nitrates and stratospheric volcanic aerosols. Even in the case of sulfur species, the gaseous and aqueous processes that convert SO₂ to sulfate and the history of SO₂ during the 19th and 20th centuries will differ among modeling groups.

Several efforts are underway to describe the parameterizations for radiative transfer used in the global modeling community. The breakout group noted two efforts underway to document parameterizations led by Q. Fu and by V. Ramaswamy under the auspices of GEWEX. Fu's talk during the plenary session on radiative forcing was based upon information assembled in a comprehensive survey of GCM radiative parameterizations that he conducted last year. During the workshop, Fu and Ramaswamy announced that they would author a joint paper on modern radiative transfer schemes used in general circulation models (GCMs). The deadline for groups that wish to include their scheme in this study is 30 September, 2004.

Benchmark forcing calculations are being assembled as part of a radiative transfer model intercomparison project (RTMIP) organized by W. Collins and V. Ramaswamy in support of the AR4. The RTMIP is soliciting calculations by well-mixed greenhouse gases from groups that

develop both GCM and line-by-line (LBL) models. The LBL results serve as the benchmark estimates of radiative forcing. Details on the experimental protocol and overall design of the intercomparison are available from <ftp://science.arm.gov/outgoing/IPCC-rt>.

Preliminary results from the intercomparison shown during the workshop indicate a wide disparity in estimates of forcing by CH_4 and N_2O at the tropopause and by CO_2 at the surface. The following spread in flux estimates was obtained: $3.45 \pm 0.98 \text{ W/m}^2$ (longwave) for the top of the model in the case of $\text{CH}_4 + \text{N}_2\text{O}$; $1.08 \pm 0.71 \text{ W/m}^2$ (longwave) and $0.85 \pm 0.76 \text{ W/m}^2$ (shortwave) for the surface in the case of CO_2 . Figure 1 summarizes some of these results.

These differences are based on preliminary results, but were viewed by most participants as surprisingly large since the calculation of radiative transfer in known conditions is a well understand aspect of climate science. Hence, the large differences presented are of great importance to understand. It seems likely that some of them will be amenable to improvement that may be very important in better understanding and narrowing the range of model forcings and perhaps climate sensitivity. It is hoped that more model groups will submit results, due to the importance of resolving these issues.

During the plenary, R. Cess also showed that the shortwave clear-sky top-of-atmosphere (TOA) fluxes from GCMs span a wide range over ocean surfaces. The breakout group recommended that the intercomparison be expanded to include as many of the GCMs submitting simulations for the AR4 as possible.

The organizers for the intercomparison suggested that the deadline for the intercomparison be extended to 15 December, 2004 to help insure participation by as many groups as

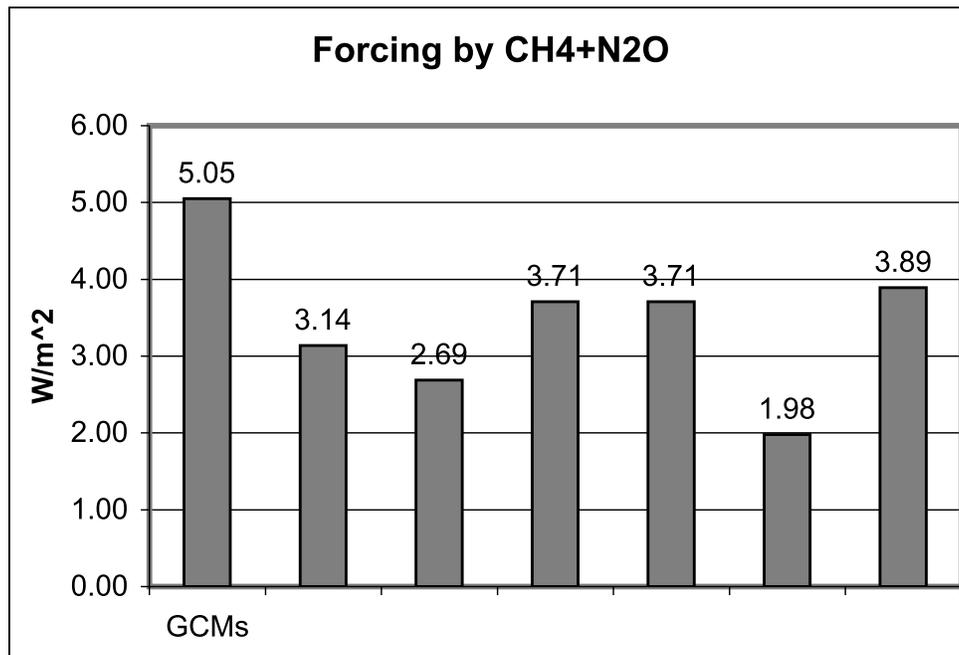


Figure 1. Total long-wave forcing by CH_4 and N_2O for the top of the model of 7 GCMs.

possible. In light of the Cess results on clear-sky fluxes, the breakout group suggested that the intercomparison also focus on absolute fluxes as well as forcings. The final results will be presented at the 85th annual meeting of the American Meteorological Society in San Diego during January 2005 and at the IPCC meeting at the University of Hawaii in March 2005. A survey of the results will be submitted for publication immediately after the IPCC meeting.

The global AEROSol model intercomparison (AEROCOM) is conducting a comprehensive evaluation of aerosol distributions, optical depths, and intrinsic optical properties against satellite, surface, and in situ observations. The data include aerosol properties from the NASA MODIS and MISR satellite instruments, the Aeronet network of scanning radiometers, and chemical speciation from surface monitoring sites. Several of the GCMs submitted to AEROCOM are similar or identical to models that will be used for simulations analyzed in the AR4. Therefore AEROCOM results could be used to document the forcing by tropospheric aerosols in some of the global models discussed in AR4. Since AEROCOM has found a wide range of global aerosol mass among GCMs and chemical transport models, the breakout group recommended that AEROCOM compare normalized forcings to eliminate disparities in mass, at least to first order. The AEROCOM group is soliciting global calculations of aerosol forcing through the end of October 2004. Results from the intercomparison will be presented at the next AEROCOM meetings in December 2004 and May 2005.

Ozone is not considered in the current RTMIP. Its history and future evolution is not specified as a part of the SRES scenarios. There was general agreement that its forcing and the treatment of ozone in GCM radiative transfer codes should be characterized in support of the AR4. Rather than expand the RTMIP to include ozone, the breakout group recommended that the ozone forcing be documented using results from the GCM Reality Intercomparison Project (GRIPS). The experimental protocol for GRIPS is described at <http://www.met.reading.ac.uk/~piers>. The group also recommended using an earlier intercomparison by Shine et al. (1995) to quantify differences in ozone forcing among global models.

The breakout group discussed the set of forcings requested as part of the archive of global model results being assembled for the AR4. The details are given at http://www-pcmdi.llnl.gov/cmip/standard_output.html. The forcings include:

1. CO₂ doubling
2. Total forcing (from all anthropogenic influences) for SRES A1B, B1, and A2 (reported at least at the end of the simulation, and as frequently as necessary to characterize, approximately, the forcing changes throughout the experiment).
3. CO₂ quadrupling (to check that logarithmic scaling holds).
4. Total anthropogenic forcing for the historical run (sampled every 10 years or more frequently)
5. Total natural forcing (if any) for the historical run (sampled as frequently as necessary)

6. Forcing due to individual constituents (e.g., volcanic aerosols, anthropogenic aerosols, individual greenhouse gases) for the historical run.
7. Forcing due to individual constituents for the experiments listed in 2.

The breakout group recommended that all modeling groups attempt to supply the first four types of forcing to facilitate analysis of their simulations.

Due to the difficulties in developing a consensus for a single method of estimating forcing, three sets of forcing calculations are requested.

- Set 1: Interactive Fluxes
 - Type: Net SW and LW fluxes
 - Levels: TOA, surface, and 200mb
 - Frequency: Monthly
 - Input: Current state information
- Set 2: Instantaneous Forcing
 - Type/level: Same as Set 1
 - Frequency: Decadal
 - Input:
 1. All species → 19th C.
 2. Individual species → current value
- Set 3: Adjusted Forcing
 - Type/level: Same as Set 2
 - Frequency: Decadal, same as Set 2
 - State information
 - Troposphere → 19th Century Control
 - Stratosphere → Adjustment via Fixed dynamical heating
 - Input:
 1. All species → 19th C.
 2. Individual species → current values

The calculations in Set 3 are equivalent to the traditional adjusted forcings discussed in the FAR. Note, however, that the new calculations adopt 200 mb as a surrogate for the tropopause rather than a more realistic definition. The calculations in Set 2 are equivalent to instantaneous (unadjusted) forcings. The calculations in Set 1 combine the effects of radiative forcing with response of the radiative fluxes to the forcing.

The breakout group made several recommendations regarding the documentation of forcing supplied with the integrations for the IPCC AR4:

1. Provision of both clear-sky and all-sky fluxes for Sets 2 and 3 would be particularly useful in separating the effects of the background atmospheric and surface state from the effects of clouds on the forcing.
2. It would be particularly helpful if the global, annual-mean TOA insolation would be included in the mandatory output fields. Inclusion of the insolation would help isolate any

differences in shortwave forcing due to differences in the solar constant, and it would help identify the reconstructions of solar luminosity used by each modeling center.

3. While the specification for the natural forcings species that the forcings should be sampled as frequently as necessary, the breakout group was concerned that volcanic eruptions might not be adequately sampled. The seasonal-mean forcing by the Mt. Pinatubo eruption should be computed continuously from 1991 through 1993, and that the seasonal-mean mass loadings should be computed for the same time period.
4. The requested fields only include mass loadings for sulfate aerosols. This will complicate comparison of normalized forcings for other important anthropogenic species, in particular black and organic carbon. The mass loadings (in kg/m^2) for black and organic carbonaceous aerosol species should be requested as well.
5. Modeling centers should document the precise methodologies used to compute the instantaneous forcings (Set 2). Some centers should conduct an intercomparison of the various methodologies using offline radiative calculations based upon the output from a subset of the models.
6. Modeling centers which include indirect effects of aerosols in their simulations should supply descriptions of their respective parameterizations for these effects and should, if possible, estimate the magnitude of these effects at periodic intervals in their integrations. At present, the requested fields do not include information sufficient to quantify the indirect effects of aerosols.

2 Breakout Group Reports

2.4 Probabilistic Measures of Climate Sensitivity: Breakout Group 4 and Climate Sensitivity and PDFs Plenary Session

Matthew Collins¹, George Boer² and Catherine Senior¹

¹ Hadley Centre for Climate Prediction and Research, Met Office, Exeter, UK.

² Canadian Centre for Climate Modelling and Analysis, Meteorological Service of Canada, University of Victoria, Victoria, Canada.

Equilibrium Climate Sensitivity (ECS)

The global equilibrium climate sensitivity, often referred to simply as the climate sensitivity, is defined as the global mean (indicated by the angular brackets) surface temperature change $\langle T'_{2x} \rangle$ experienced by the climate system, or a model of the system, after it has attained a new equilibrium after a doubling of atmospheric CO₂ concentration. More generally, the global equilibrium climate sensitivity parameter, s , is the link between the CO₂ (or other) perturbation to the radiation stream, measured by the radiative forcing $\langle f'_{2x} \rangle$, and the resulting temperature response in the form

$$\langle T'_{2x} \rangle = s \langle f'_{2x} \rangle.$$

The utility of s extends beyond its global-mean value, containing information on the geographic pattern of temperature response for different forcing agents (e.g. Boer and Yu, 2003).

Although (we suspect) s is a single deterministic number, we do not know its precise value. This uncertainty in the value of s , or any other measure of climate response, can be formally characterised by a probability density function, $p(s)$, where the PDF in this case represents our uncertainty in the actual value of the parameter. We use s and $p(s)$ to represent the ECS and its PDF in what follows for notational convenience, with the understanding that s can be replaced by the global mean temperature measure or any other metric of climate response.

PDFs of Climate Sensitivity

Probabilistic approaches aim to quantify the uncertainty in s by producing PDFs of possible values. PDFs are of use to both the scientific and the policy communities as they assign likelihood to relatively high or low sensitivities rather than simply stating a range (as in previous IPCC Assessments) with no estimate of the probability of a particular sensitivity actually applying to the climate system. Rather than a range or a “best guess”, $p(s)$ is used to formally indicate the probability that s falls within a particular range. In order to more clearly delineate the uncertainty in the value of s and to permit its unambiguous reduction, its uncertainty must first be quantified in terms of $p(s)$.

Since the publication of the TAR, a number of studies have produced PDFs of s and related metrics. All studies employ models and observations of the climate system and use ensemble/

statistical techniques to compute probabilities. The basic approach is characterised by Bayes' Theorem, namely

$$p(s|\text{data}) \propto p(\text{data}|s)p(s),$$

in which the *posterior* probability of s constrained by the data (model and observations) is proportional to the *likelihood* of the data assuming a given s (the model-observation fit) multiplied by the *prior* probability of s . Ideally $p(\text{data} | s)$ is so well constrained (a narrow PDF) that the choice of prior is unimportant, but this is not the case here.

The models used are of a varying level of complexity, the observations are of varying quality (in the sense of the inherent error characteristics) and the ensemble/statistical techniques are different. Moreover, the sources of uncertainty that are quantified are different. Studies typically use a range of *constraints* on s and are *conditional* upon a range of different assumptions in their construction. Each study produces a $p(s | \text{data})$ which reflects the uncertainties assigned to the data and model from which it is derived and these PDFs therefore differ among themselves. As noted by Frame and Allen (2004), the result may depend critically on the choice of the prior, particularly the high-sensitivity tail of the distribution.

General Method

The general approach can be characterised as follows:

1. Observations which on physical grounds are thought to constrain s are gathered over a particular time period.
2. An ensemble of model experiments is performed for the time period in question in which models and/or their parameters and forcings are varied. (In the simplest case the “model” condenses to a formula.)
3. Ensemble members are given relative weight depending on their ability to reproduce the observations. (This step is sometimes omitted, together with step 1, and all ensemble members are given equal weight.)
4. The PDF is extracted from information contained in the ensemble.

Details of individual studies which have been published, which are likely to be published on the time scale of the AR4, or which were presented at the meeting can be found in the appendix of this report.

General Features of PDFs

The majority of PDFs produced have a skewed distribution with a relatively sharp “cut-off” at low values, a modal value less than the median and a “fat-tail” extending to high values

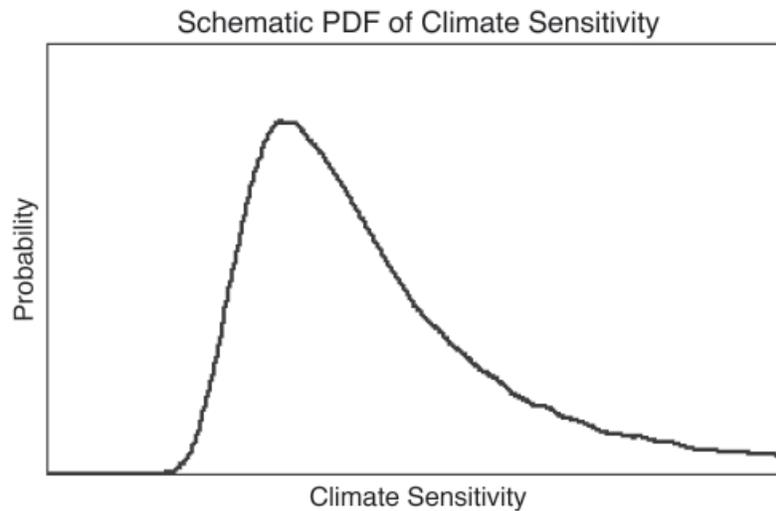


Figure 1. A schematic figure highlighting the general shape of the PDF of climate sensitivity, as found in a number of studies.

(Figure 1). The existence of this fat-tail is of particular relevance for policy as it is related to the probability of “dangerous” climate change. Several explanations have been put forward for this:

1. Uncertainties in feedbacks (clouds, sea-ice etc.) tend to be normally distributed and the shape is due to s being proportional to the inverse of the feedback parameter.
2. Measures of model-data fit used to give relative weight to ensemble members tend to produce distributions with fat-tails (e.g. F or χ^2).
3. There is a non-linear relationship between the s and observed warming. High sensitivity can be countered by high ocean heat uptake to reproduce the observed trend in global mean temperature.

All these explanations are plausible and it seems that the shapes of the published distributions can be explained by one (or more) of the reasons.

Summary

Considerable progress has been made since the TAR in attaching estimates of the probability of different values of s and related metrics. A number of independent methods have been developed which employ a wide range of climate models and a wide range of different observations. No method has however emerged as being superior to another. The “range” of uncertainty in s may now be more clearly defined in probabilistic terms in the sense that, for a particular PDF, the probability of the equilibrium climate sensitivity being less than 1.5 or greater than 4.5 (or any other range) may be stated. The inherent uncertainty in observations, forcings and climate models means that the PDF of s and other commonly used metrics remains relatively broad.

There is a considerable challenge to present probabilistic information in a clear and useful way. Because the different studies all account for different uncertainties, use different models and different observations and have different prior assumptions, the combination of information from different sources into a single PDF is not currently possible. However, the individual PDFs may be grouped, put in context and annotated in a “multi-PDF diagram”. This is the suggested approach for the AR4.

Report from the Breakout Group

There was a wide-ranging discussion during the breakout group under the headings of “Estimates”, “Synthesis” and “Utility”. The following is a summary of the key points and recommendations.

- Methods which use observed historical changes require estimates of the radiative forcing over the period. These contain large uncertainties, particularly in the direct and indirect forcing from aerosols. It was recommended that GCM groups submit forcing fields to PCMDI.
- PDFs from different sources cannot be combined because of the different priors and the different sources of uncertainty considered. The “multi-PDF plot” was recommended as a format for showing PDFs in the AR4 (e.g. grouped by prior). There should also be a commentary on why the PDFs are different.
- It was agreed that the “perturbed physics” methodology was useful and that the method proposed by Senior et al. (2004) – known as the Wielicki proposal - to relate model sensitivity to observed variables should be pursued. Only limited results could be available for the AR4 though.
- The multi-model approach has utility, although we should be aware of the “caveat of convergence” whereby all modelling groups would (unintentionally) tend not to produce models which appear to be outliers.
- There was some debate over the use of measures of model quality (e.g. the Climate Prediction Index of Murphy et al. (2004)) and some who thought that those GCMs which have very high or very low sensitivities should be subjected to an open expert assessment.
- The Equilibrium Climate Sensitivity (ECS) is a familiar metric and hence is still policy-relevant. It was agreed that the subject has advanced enough such that it will be possible to avoid re-stating the simplistic range in the AR4. The addition of other information e.g. Transient Climate Response (TCR) and plots of temperature change with time is also advisable.
- There was some hope that some “categorical” statements, particularly regarding the low-sensitivity tail and TCR/future warming may be made. Here categorical refers to the form of statements suggested by Moss and Schneider (2000).

Appendix: A Summary of Studies

The following table highlights the main features of studies which have been published or are likely to be published on the time scale of the AR4 or which were presented at the meeting. All authors were cognisant of the uncertainties in their estimates and hence presented a number of different PDFs in their papers. It is not therefore possible to give, for example, a median and 5–95% range for each study.

Study	Metric evaluated	Model(s) used	Observations used	Comments
Andronova and Schlesinger (2000)	ECS	EBM	Historical warming and radiative forcing	
Wigley and Raper (2001)	1990 to 2100 warming	EBM	Historical warming and radiative forcing	Many different scenarios considered
Forest et al. (2002)	ECS	EMIC	Historical surface, atmospheric and ocean temperature changes and historical radiative forcings	
Knutti et al. (2002)	2100 warming under SRES B1 and A2	EMIC	Historical surface and ocean temperature changes and historical radiative forcings	
Gregory et al. (2002)	ECS	SF	Historical surface and ocean temperature changes and historical radiative forcings	
Stott and Kettleborough (2002)	Transient 21 st Century warming	AOGCM	Historical surface temperature changes	Examine 4 SRES scenarios
Forster and Collins (2004)	Water vapour feedback	AOGCM	Surface temperature and water vapour changes following the Pinatubo eruption	
Murphy et al. (2004)	ECS	AOGCM	“Basket” of many different time-mean observations of climate	Parameter perturbations to a single atmosphere-slab model
Stainforth et al. (submitted)	ECS	AOGCM	6 time-mean surface variables	Very large ensemble of parameter-perturbed atmos-slab models
Hegerl et al. (extended abstract)	ECS	EBM	Palaeo (last 1000 years) reconstructions of surface temperature variations and radiative forcings	
Frame and Allen (extended abstract)	ECS and TCR	EBM	Historical surface and ocean temperature changes and historical radiative forcings	Note the importance of the prior
Räisänen (extended abstract)	ECS and TCR	AOGCMs	None	Multi-model estimates from TAR models
Von Deimling et al. (extended abstract)	ECS	EMIC	Last Glacial Maximum tropical SST changes	
Tebaldi et al. (2004)	Warming under SRES A2	AOGCMs		

A note on models

A wide range of climate system models have been used to produce PDFs. For the purpose of this note, models have been stratified as below and their properties which are particularly relevant to the production of PDFs noted.

- *SF – Simple Formula*: In at least one study (Gregory et al. (2002)) a simple linear formula which relates changes in global mean temperature to changes in a flux of energy has been used.
- *EBMs – Energy Balance Models*: These are the simplest climate models which are cheap to run and easy to understand. The ECS is in virtually all cases a prescribed parameter in the model.
- *EMICs – Earth System Models of Intermediate Complexity*: These are more complex models than EBMs. The ECS is generally not a prescribed parameter in the model, however they are usually simple enough and fast enough to run that complete smooth mapping between the model parameters and the ECS can be established.
- *AOGCMs – Atmosphere Ocean Global Circulation Models*: These are the most expensive to run and most complicated to understand. The ECS is a complex (perhaps non-linear) function of the model parameters.

While AOGCMs are the most complex models, they do have the potential for making more impact-relevant probabilistic predictions of regional changes, changes in variables other than surface temperature and changes in extreme events.

3 Overview and Key Conclusions Plenary Sessions

3.1 Introduction and Overview

Rapporteur: Caspar Ammann

Review of concepts and methods relating to climate sensitivity

Jonathan Gregory (Univ. Reading & UK Met Office – Hadley Centre)

Climate Sensitivity is a useful term because many aspects in a climate model (spatial and seasonal patterns of change) scale well with global average temperature (ΔT). ΔT is a fairly well known measure in the real world. Such scaling properties are generally true in any given model, but not necessarily across models. However, it does not include variability.

It is helpful to distinguish forcing and feedbacks. Effective forcing is a radiative response to a forcing agent (*instantaneous* radiative flux changes as well as *induced* such as stratospheric adjustment), while feedbacks are a response to climate change. Climate feedbacks can be dependent on the forcing (magnitude and overall nature). Shine et al. (2003) show different response to various instantaneous forcings. However, if full feedbacks (tropospheric and stratospheric climate adjusted) are allowed, then the climate response ($\Delta T/Q$) is very similar.

Sensitivity is not constant from a transient viewpoint. The fast feedbacks scale well with ΔT . Slow feedbacks, however, not only have their own timescale (THC, vegetation), they are acting more as a forcing rather than a direct feedback (changing climates are influencing SSTs and thus patterns of cloud feedback). Consequently, coupled transient simulations show changes in slope of top-of-atmosphere (TOA) radiation versus ΔT . Therefore, determination of *Climate Sensitivity* has to be seen as the sum of processes that cause radiative responses, including the response from feedbacks where local effects end up determining global processes and thus to a large part the sensitivity (Boer and Yu, 2003).

In time-dependent climate change, the global temperature response ΔT is determined by both climate sensitivity and the ocean heat uptake. A general anti-correlation of heat uptake and sensitivity would reduce the range of transient climate response.

In recent studies, the use of observed variability and change to constrain local and global feedbacks has complemented the evaluation of climate sensitivity from GCMs.

Understanding and comparing climate sensitivity and feedbacks in models

Robert Colman (BMRC, Melbourne, Australia)

Quantification of forcing and sensitivity in a model requires an understanding of the feedbacks. This includes the critical water vapor feedbacks, for which simulations to $2\times\text{CO}_2$ have shown increases in upper troposphere water vapor of +66% in the tropics and +33% in higher latitudes. These changes show their effects in different ways with regard to the visible versus the infrared

spectrum causing an asymmetric response in the annual cycle. In LW spectrum, water vapor feedbacks change ~10% over the annual cycle (little changes in clouds and lapse rate), but SW annual cycle shows enormous variations mostly from cloud changes. In fact, the cloud feedback (albedo) actually changes sign over the annual cycle. Therefore, the annual cycle needs to be resolved.

Overall, lapse rate changes can be offset by the water vapor signal (at least in BMRC model but not in GFDL, Soden et al.). Model intercomparisons including Radiative Convective Models seem to confirm offsetting effects. This would reduce the dependence on lapse rate changes. But how constant is the relative humidity? In Pinatubo experiments of Soden et al. (2002) a very stable rH was consistent with observations.

Interpretation of model results that show changes of climate sensitivity with time

Sarah Raper (AWI, Bremerhaven, Germany)

Conventionally, climate sensitivity is defined as the equilibrium warming for a forcing at double CO₂ concentrations in a model using a slab/mixed layer ocean. Because these models are not the same as the ones used for transient climate change simulations, the resulting climate might be different. Boer and Yu (2003) show the connection between feedbacks and sensitivity.

Murphy (1995) defined the effective climate sensitivity as equilibrated temperature change at doubled CO₂ with feedbacks held fixed. Such a measure is more of a hypothetical concept than a directly measurable quantity in climate models.

Is the sensitivity indeed constant? Critical in calculation of sensitivity are heat exchange coefficients of land/ocean with the atmosphere. Feedback parameters lambda for land and ocean are required to satisfy the given climate sensitivity. The exact ratio between the two is unclear. Using MAGICC, three different (extreme) exchange coefficient combinations are evaluated: 0.5, 100 and 16 (with 16 corresponding to 3.6 W/m² for a warming of 1.6 K in the PCM model). If the sensitivity is an input to the model, then the effective climate sensitivity for the three exchange coefficients is found to be changing. Only in a very special case when land and ocean changes are relatively fixed is the sensitivity actually constant. More common is that global mean temperature changes result in changes in patterns of response. This leads to a change in sensitivity over time. Thus, it is suggested that for the AR4 a more flexible measure of climate sensitivity should be used. For example, the transient response (ΔT at 2xCO₂ in 1% experiment) would be a more transparent and relevant metric for model comparison.

Probabilistic estimates of climate sensitivity from GCM ensembles

James Murphy (UK Met Office, Hadley Centre)

The current state of affairs allows ensemble predictions using a small number of simulations without information on how reliable individual results might be. One constraint generally applied

is the fit of the simulation to the historical record. But uncertainty arises similarly from initial conditions, ensemble size, and modeling uncertainties (parameterizations).

An effort to systematically evaluate the real range of uncertainty in HadAM3 with mixed layer ocean is performed using a ‘perturbed physics ensemble’. Initially, a selected set of 29 ‘most important’ parameters is varied. The experiments are run for $2\times\text{CO}_2$ and evaluated in the measures most closely related to sensitivity (such as SW cloud forcing). The resulting sensitivity ranges from 2.25 K up to about 4.25 K, although one realization exhibited a change of 7 K. The mean was close to the standard model.

But how can one go from frequencies to probabilities? Comprehensive sampling is currently not possible since only single parameters are perturbed, not combinations. In later stages such combined variations will be performed.

To quantify a reliability of such simulations, a Climate Prediction Index (CPI) is established based on the deviations from multi-annual mean fields of present day climate variables. Currently, an average RMS of 5 is found for the CPI, representing clear biases in the model. Most uncertainty results from the cloud occurrence and related radiative forcing.

Using linear statistics, estimated probabilities for climate sensitivity can be derived (see Murphy et al., 2004, Nature). The straight forward un-weighted PDF (all model version equally likely) based on physical understanding of parameters as a constraint gives a mean sensitivity of about 2 K. If however the simulations are weighted by the fit to present day observations, then the mean sensitivity is closer to 3.5 (95%: 5.4 degrees). Multi-parameter perturbation simulations are ongoing. Linear predictions of how the multi-parameter results might look like seem to underestimate the sensitivity (preliminary results!).

Climate Sensitivity inferred from PaleoRecords

Jean Jouzel (IPSL/ LSCE Paris, France)

“Our most precise knowledge of climate sensitivity comes from data on ancient and recent climate change.” (Hansen et al., 1993). However, in order to better use past climate information, it is required that we improve our knowledge of climate forcings, reduce uncertainty in climate response evaluation (from a large variety of indicators), and that the strong interaction between paleo climatologists and modelers are further expanded. Additionally, it is important to recognize that there is no satisfactory past analog for future climate change.

Deep time (Mid-Cretaceous, early Eocene): $\Delta T_{2\times\text{CO}_2} = 2-5 \text{ K}$

However, critics (e.g., Lindzen) argue that differences in heat fluxes due to changed latitudinal gradients have to be considered. Other criticism (Shaviv and Veizer, 2003: $\Delta T_{2\times\text{CO}_2} = 0.75 \text{ K}$) does not seem to withstand detailed tests (Rahmstorf et al., 2004).

Last glacial periods: $\Delta T_{2\times CO_2} = 3-4 K$

Well documented at global scale and relatively stable climate over longer periods allows establishment of equilibrium climate. The estimates of Greenhouse forcing for last glacial $\sim 2.3 \text{ W/m}^2$ (Lorius et al, 1990). Other factors such as dust, sulfate and local insolation changes have to be included. 40–65% of temperature change due to greenhouse forcing. Other estimates in same range (Hansen et al., 1993; Hoffert and Covey, 1992). New ice core (EPICA Dome C) now covers eight glacial cycles, maybe nine (EPICA, 2004). There is a clear change of pacing before 400k, a fact that requires resolution.

Last Millennium: $\Delta T_{2\times CO_2} = 2-4 K$

Climate response to solar and volcanic forcing from ice cores used in Energy Balance Models and first GCM simulations show good agreement with proxy reconstructions. Although the magnitude of solar irradiance changes is unknown, large values seem to cause climates inconsistent with climate reconstructions. Additionally, changes in atmospheric CO_2 can help to estimate the magnitude of climate change and thus sensitivity.

How can we constrain the cloud-radiative feedback produced by climate models?

Sandrine Bony (LMD/IPSL, Paris, France)

Clouds strongly affect the Earth's radiation budget. Currently, the net forcing (LW warming vs. SW albedo cooling) is negative, but how will this change with future warming? (see Bony et al., 2004). Use of IPSL-CM4 GCM to illustrate changes in CRF in transient 1% CO_2 run with 2.2 K for $2\times CO_2$, and 4.4 K at $4\times CO_2$. Currently 20 W/m^2 net cooling from clouds, but cooling effect decreases with time into the future by several W/m^2 as cloudiness decreases. However, there is no clear consensus between models at this point (see difference GFDL and NCAR models, B. Soden).

Question is what should be diagnosed to constrain the CRF on climate change time scales? With only ~ 20 years of observations, this is not an easy task. Here one approach is described that focuses on cloud changes in tropical regions using large scale circulation regimes with changes in area between ascending and descending motion (e.g. using OMEGA as a parameter).

The IPSL GCM shows a weakening of the Hadley circulation with increasing CO_2 . However, what dominates the change in tropically-average CRF is the change in cloudiness that occurs in subsidence areas. But how can we constrain the behavior of clouds in these circulation regimes? E.g., relationship of clouds to SST can be used, but how relevant is this short-term relationship (with natural variability being dominant) for changes in the long-term climate change? It is unclear how robust estimates are that use the short observations, even with regard to their sign.

Using circulation regimes might give some guidance on model evaluation. Comparing, for different circulation regimes, the SST dependence of cloudiness computed by climate models with that derived from observations (by using satellite data together with NCEP or ERA40

reanalysis) might help to constrain the global cloud radiative feedback. In that analysis, particular attention should be put on regimes of moderate subsidence, covered mostly by cumulus and stratocumulus clouds, because of their large statistical weight on Earth. On longer terms, improving our physical understanding of these feedbacks will help to define better observational constraints. On longer terms, it is required that we improve the physical understanding of these feedbacks.

Summary

Climate Sensitivity defined originally as equilibrium global temperature response to radiative forcing ($2\times\text{CO}_2$) is useful because many climate properties scale well with global temperature. However, single numbers of surface temperature changes do not seem to be ideal descriptors from a transient climate change point of view because they are dependent on feedbacks and ocean heat uptake. Additionally, effective climate sensitivity in a given model is not a constant number but appears to be changing depending on forcing and climate state. Evolving climate sensitivity, e.g. temperature change at $2\times\text{CO}_2$ in a transient experiment, is suggested as more transparent measure for model comparison in AR4. Additionally, sensitivity at the global scale might be combined with a measure of variability to more comprehensively capture the changes in the climate system.

Better understanding of climate sensitivity is dependent on improved knowledge about critical feedbacks in the climate system. Model intercomparisons are required to study feedbacks such as water vapor and lapse rate changes. It is important to resolve the annual cycle because certain feedbacks don't apply in a uniform manner throughout the year. Changes in spatial patterns of cloud response over time represent a significant source for change of transient climate sensitivity. Currently, detailed analyses of observational and model data are still in a conceptual stage but could move towards more dynamically motivated intercomparisons, e.g. using distinct circulation regimes to separate cloud response.

Uncertainty related to observational data and models remains considerable. A 'perturbed physics ensemble' approach using at first single and later multi-parameter perturbations can better pinpoint uncertainty in climate models and quantify their influence on measures like global climate sensitivity. Probabilistic approaches are more appropriate than single numbers for individual climate models.

Past climates offer some guidance to climate sensitivity. Estimates generally fall in a range of equilibrium temperature change of 2–4 K (for $2\times\text{CO}_2$). However, it needs to be recognized that there is no clear analogue to future climate change in the Earth's past. This is particularly the case if climate sensitivity is dependent on the climate state as well as character of the forcing. Therefore, comparison with past climate changes have to be performed with caution.

3 Overview and Key Conclusions Plenary Sessions

3.2 Key Conclusions

Rapporteur: Tahl Kestin

The last three speakers in the workshop received a special challenge from the workshop organizers: to present their personal view on the key results that have emerged from the workshop, and their implications.

Richard Somerville is one of the Coordinating Lead Authors for Chapter 1 of the WG1 AR4, “Historical Overview of Climate Change Science”. His talk and the associated discussion suggested three key results with respect to clouds and radiative transfer that have emerged from the workshop:

1. *Cloud effects still dominate climate sensitivity, but other processes also merit attention.* The intercomparison of models with the same atmosphere or the same ocean (Gerald Meehl) reinforces what has been known for a long time: that atmospheric models, and within them the cloud components, largely determine the climate sensitivity of a model. However, recent work by Colman, Hall, and others as discussed during the workshop highlights the importance of water vapor, lapse rate, and planetary albedo as significant factors as well.
2. *Radiative transfer in models needs improvement.* Intercomparisons of GCM radiation codes and line-by-line (LBL) models (William Collins) show that GCM results have a large variability, in part because some omit important gases at both the short-wave and the long-wave. Greater attention also needs to be paid to radiative transfer methods and to surface versus top-of-atmosphere flux differences, which are surprisingly large. These are problems that should be fixable in time for the AR4.
3. *Aerosol effects are especially poorly understood.* Aerosols have many properties and interactions (Stefan Kinne), many of which are not well represented in models. Consequently, aerosols can have either a cooling or warming effect on local scales. Therefore, although the radiative forcing of aerosols is much smaller than that of GHGs for global scales, it can be much larger on local scales; we should highlight this difference.

These results suggest that:

1. *Cloud parametrizations require attention to comparisons with observations.* Cloud radiative processes are determined by both cloud macrophysics and cloud microphysics. Although some models include a few cloud microphysics parameters (James Murphy), none reproduce the ISCCP cloud climatology (Bryant McAvaney), and few agree on even the simplest aspects of cloud changes with global warming (Sandrine Bony).
2. *To be acceptable to IPCC, models must pass tests.* Models can be tested in several ways—for example in their ability to reproduce the collapse of the Walker circulation during El Niño (Robert Cess), and the temperature response to the Mt. Pinatubo eruption (Alan

Robock and Piers Forster). They should also get the right answers for the right reasons. Although there is sensitivity to testing models, it could be done objectively (for example by assigning weights or acknowledging the shortcomings of models) and would enhance the influence and credibility of the AR4.

3. *Climate sensitivity cannot be only one global number.* Climate sensitivity is a convenient measure for a particular model, but it differs from model to model and is not constant (Jonathan Gregory). It focuses on equilibrium climate and does not capture climate variability and regime changes. It is a global mean quantity and does not reflect the possibly large local effects of clouds (as with aerosols). And it may mask large changes in the annual cycle of cloud feedbacks (Robert Colman).

Thomas Stocker is one of the Coordinating Lead Authors of Chapter 10 of the WG1 AR4, “Climate Model Projections”. His talk reflected on the key conclusions of the workshop with respect to observations and paleoclimate. In order to estimate “relatively simple” metrics of climate sensitivity, we need to understand the complexity of and effectively use a hierarchy of models and a hierarchy of time scales. However, for all time scales we need to better understand the uncertainty in the forcing estimates.

- *The last 20 years:* The availability of global, 4-dimensional, and high-resolution data allows many kinds of analyses and tests of models, including examining the seasonal response, extending the tests of models with the Mt. Pinatubo eruption (Soden et al., 2002) to other OAGCMs, using satellite data and reanalysis products to estimate individual sensitivities, and using new ocean and tracer data to constrain the transient climate response and the change in forcing.
- *Last 150 years:* Instrumental data could be combined with ensemble simulations. Additional types of data, such as anthropogenic tracers could be used to constrain climate sensitivity.
- *Last 500 years:* Reconstructions now provide numbers with units that can be input directly into paleo-models to provide seasonal and regional temperature changes. These reconstructions are currently available mainly for Europe, so there is need to extend them to North America and Asia.
- *Last 1000 years:* A major effort is needed to reconcile the various Northern Hemisphere temperature data sets that were compiled by Briffa and Osborn (2002), for example by reducing the uncertainties through model simulations. Further, there is need to extend these temperature data sets to areas of the Southern Hemisphere with enhanced response (e.g., Antarctica). There are also opportunities to use ice-core proxies to estimate SST changes outside Antarctica, and to use biogeochemical data to constrain changes in vegetation and land use.
- *Last Glacial Maximum (LGM):* Both comprehensive and simplified models can be used to understand LGM temperature changes, and ensembles and model intercomparison projects

can help to account for the uncertainties in these models and the data. Further, there is need to study the potential of translating regional temperature changes (such as in tropical temperatures) into global climate sensitivity.

- *Previous interglacials*: The new Dome Concordia 740,000-year data set of temperature and forcing (through dust load), adds four older cycles to the four known recent cycles. The newly-identified older cycles have warmer peaks than those of the recent cycles. Challenge: can modelers predict the change in CO₂ that will be found in the ice cores to be associated with the warmer peaks?

Given the available data, the needs, and the opportunities, the following research priorities that would most benefit the AR4 were suggested:

1. *The last 1000 years*: Reconcile the Northern Hemisphere temperature records.
2. *Last 20 years*: Understand the complexity of the climate signal through top of atmosphere and ocean data, and responses to Mt. Pinatubo and ENSO.
3. *Last Glacial Maximum and previous glacial/interglacials*: Better understand the relationship between CO₂ and warm periods, and better understand tropical responses.

The discussion of the presentation focused on the need to reduce and be honest about the uncertainties in the paleo-observations, and to take advantage of more recent data, which has smaller uncertainties (i.e., that the last 150 years should be a focus in the last 1000 years, and consider its implications for the uncertainties in the last millenium among the list of priorities).

David Randall, who is one of the Coordinating Lead Authors of “Chapter 8: Climate Models and their Evaluation”, in the WG1 AR4, spoke on behalf of Chapter 8 on the implications of the discussions at the workshop for the chapter. The alternative title for his presentation was “A plea from Chapter 8”.

One of the hardest and most important questions that Chapter 8 needs to answer is Question 6 from BOG1: “Why have your results changed?”, or why have the estimates of climate sensitivity from the models changed from the TAR estimates?

The key to answering this question and to bolstering the credibility of the answer is to view models as a collection of concepts, rather than just software (see Figure 1 on next page), and to try and explain the changes in the models (and hence climate sensitivity) in terms of conceptual changes in model design. Furthermore, these concepts should be tested through process-level tests against observations. Randall placed particular emphasis on the importance of cloud-resolving models capable of examination of the full range of cloud processes, and the need to better understand the limitations of cloud parameterizations in GCMs based upon such studies (including LES, or Large Eddy Simulation).

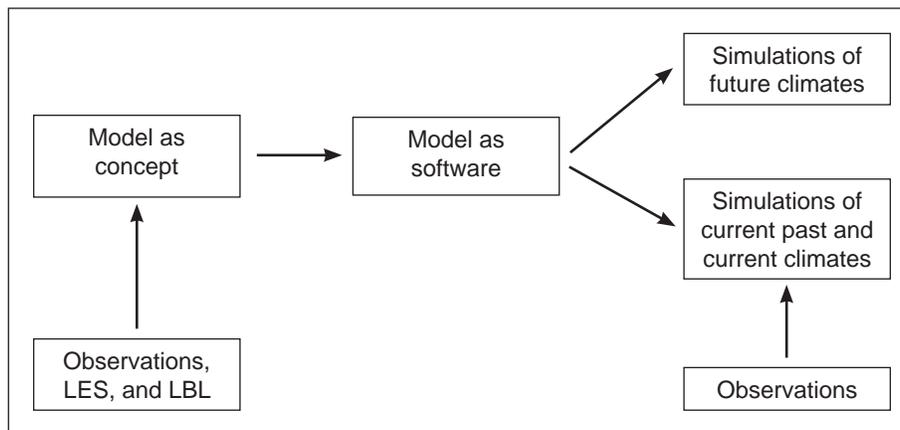


Figure 1. Schematic view of models as a collection of concepts.

Therefore, for Chapter 8, BOG1 Question 6 needs to be extended from “Why have your results changed?” to:

- What changes have you made in your model?
- What were the rationales for these changes?
- What changes did you *expect* to see in the results?
- What changes did you *actually* see in the results?

And perhaps:

- Can you provide information regarding the confidence in your cloud parameterizations?

4 Thematic Plenary Sessions

4.1 Climate Sensitivity from Models

Rapporteur: Richard P Allan

Jeff Kiehl (NCAR) reported on comparisons between the GFDL and NCAR model climate sensitivity using a hierarchy of model configurations including AMIP-type, slab-ocean and fully coupled atmosphere-ocean models.

Comparing AMIP versions of the models showed close agreement with each other and the observations. This included, for example, spatial scatter of cloud radiative effect against reanalysis vertical motion fields as well as the correspondence of regional precipitation anomalies with ENSO index. Using developmental versions of the slab-ocean models, large changes in climate sensitivity were reported. Despite similar current surface temperature sensitivity to CO₂ doubling (T_{2x}) of about 2.9 K, there are large regional differences, for example in the response of marine stratocumulus to climate warming. Examining coupled 1%/yr CO₂ increase experiments, a larger sensitivity and interannual variability of surface temperature was present in the GFDL model. In these experiments the GFDL model produced increases in total cloudiness while the NCAR model produced reductions. It was argued that structured SST change experiments offered a better means of comparing processes in the models (for example, how do low clouds respond to a particular $2\times\text{CO}_2$ -like SST change imposed on each model?).

An important point was raised with regard to 20th century climate simulations. It was argued that the agreement between these simulations and the observed temperature changes is overstated because models with larger radiative forcing in fact tend to also have smaller climate sensitivity. The spread in radiative forcing between models, mostly relating to differences in aerosol forcing, is larger than expected and should be reduced.

Herve Le Treut (LMD) talked on dynamical feedbacks. A change in the relative sizes of convective and subsidence regions may also contribute to climate feedbacks. In a simple model a 10% increase in the moist area was found to produce a 1.5 K warming. It was noted that some GCMs produce a reduction in the area of convective regimes. Other potential processes included non-linear responses of mid-latitude energy transport to sea surface temperature (SST). Simple and controlled intercomparison experiments were called for to test some of these possibilities.

Bryant McAvaney (BMRC) noted the sensitivity of shortwave cloud radiative forcing response to the radiation scheme used for Cess-type ± 2 K SST experiments. Preliminary results from the CFMIP project showed large differences in ISCCP diagnostic clouds simulated by the models compared to each other and the ISCCP observations. For example, thin mid and low-level clouds appear underestimated by all models considered compared to ISCCP. Cluster analyses that can objectively identify cloud types, for example thick, high cirrus or shallow convection, were presented as a function of cloud top pressure or total water path and optical depth. 1997/8 minus 1987/8 ENSO experiments were performed; the LMD model could predict the circulation and cloud changes but the BMRC model could not.

Ron Stouffer (GFDL) described comparisons between coupled and slab-ocean $2\times\text{CO}_2$ experiments with the GFDL R15 model. One of the main motivations for this was the question, is the slab-ocean model a good surrogate for fully coupled climate model prediction? If this was the case, there is little need to run the computationally expensive fully coupled models. The results suggested that the patterns of warming and precipitation changes were similar in both experiments but that the magnitude of the warming was larger in the coupled experiment ($T_{2\times}=4.5^\circ\text{C}$) than for the slab model ($T_{2\times}=3.7^\circ\text{C}$). It was also noted that anomalous cooling over the southern Indian ocean in the coupled version were related to spatial changes in the ocean subduction zones. It is not obvious that these results apply to other models however.

Anthony Broccoli (Rutgers) talked on climate sensitivity as diagnosed from developmental versions of the GFDL AM2 model. Climate sensitivity for the model versions, calculated using the partial radiative perturbation (PRP) technique, ranged from about 2.5 to 4.5 K. One of the biggest changes in climate sensitivity arose on updating the planetary boundary layer scheme. Consistent with previous work, there was compensation between water vapour and temperature lapse rate feedback. The range of lapse-rate feedback was comparable to the range from cloud feedback that in turn was determined primarily by the range in the shortwave radiative component. Water vapour, albedo and longwave cloud feedback ranges were relatively small. Compared to other models, the final AM2 version exhibits a strong positive water vapour feedback that is compensated by a strong negative lapse rate feedback. Note that this is consistent with a large upper tropospheric temperature response to surface temperature and constant relative humidity.

The PRP method was compared with the standard Cess-type $\pm 2\text{K}$ SST experiment method in determining feedbacks. Significant differences in cloud feedbacks were found with the Cess method giving smaller net cloud feedbacks than the PRP method – this relates to the unrealistic nature of uniform SST changes compared to CO_2 induced warming with the pattern of SST change influencing the model feedbacks. Prescribing structured SST changes in the Cess method gives results closer to the PRP method. It was also noted that changes in cloud radiative forcing are not equivalent to cloud feedbacks due to the masking of non-cloud feedbacks. For example, negative changes in net cloud radiative effect do not necessarily denote negative cloud feedback. The differences between the two diagnostics are, however, fairly easy to correct for.

Gerald Meehl (NCAR) compared feedbacks and ocean heat uptake in 5 models, varying either the atmosphere or the ocean/sea-ice components and performing 1%/yr CO_2 increase transient climate experiments. Changing the ocean did not impact the calculated climate sensitivity as much as changing the atmospheric models. A colder base state was shown not to necessarily produce a bigger surface albedo feedback. The model with the largest ocean heat uptake corresponded with the lowest climate sensitivity. However, the ocean heat uptake is dependent on a number of competing mechanisms and is model dependent. It was found that the meridional overturning of the ocean was highly sensitive to the atmosphere model. This is explained by the different regional atmospheric feedbacks altering the surface water freshness, for example by evaporative feedbacks, which impinge upon the density driven ocean currents. However, the two most sensitive models disagreed on the sign of change in meridional overturning. While the equilibrium climate sensitivity is instructive, it is necessary to understand the transient

response. Model-dependent ocean heat uptake processes must be taken into account, along with atmospheric feedbacks, to understand the transient climate response.

Akio Kitoh (MRI) reported on recent changes in two Japanese models' climate sensitivity, which were the extremes in the TAR. The updated version of the MRI model contained a more realistic simulation of low cloud radiative effect and a smaller change in shortwave cloud forcing response to CO₂ doubling. This increased the climate sensitivity from about 2 to over 3 K. A reasonable simulation of 20th century temperature was reported although rapid warming in the 1970s is earlier than the observed rapid warming of the 1980–1990s. There is a resolution dependence of precipitation with decreased convective and increased large-scale precipitation with increased resolution for the MRI CGCM3.

The MIROC3.2 model, run on the Earth Simulator, was analysed. Two versions of the model were presented, one with a lower climate sensitivity of 4 K and the other with a climate sensitivity greater than 6 K. The reasons for the different sensitivity relate to the treatment of cloud water phase diagnostic and the treatment of melted cloud ice. The climate change experiments produce a large increase in shortwave cloud forcing (more heating) which explains the high climate sensitivity. Although the present day cloud radiative forcing for the latest models looks reasonable in comparison with ERBE, the original version of the high sensitivity model produced a cooling rather than heating in response to the 20th century forcings. This was found to be due to an unrealistically large indirect aerosol effect. The updated higher sensitivity model now agrees better with 20th century observed temperatures although it still produces too small a warming from 1930–1980. This may relate to an unrealistically large response to volcanic forcing. Presumably this should be straightforward to test against the observations of the Pinatubo eruption. Finally, last glacial maximum experiments suggest that all models are within the uncertainty of the paleoclimate proxy data.

4 Thematic Plenary Sessions

4.2 Paleoclimatic Observations and Climate Sensitivity

Rapporteur: V. Masson-Delmotte

The five presentations of this section were dedicated to different time scales of past climate change relevant to different aspects of climate sensitivity: the last millennium and the carbon cycle-climate feedbacks; the mid-Holocene climate change with particular aspects of seasonal and hydrological cycle changes; the last interglacial period and the temperature thresholds associated with a possible melt of the Greenland ice sheet; and the last glacial maximum climate change.

The presentation of **Fortunat Joos** was focused on the last millennium, a period for which the natural radiative forcings remain poorly quantified. Sensitivity studies to three different solar forcing scalings from cosmogenic isotopes were performed with the NCAR model from year 900 to year 2000. The comparison between model results, 20th century trends and northern hemisphere temperature reconstructions rules out the small and large solar forcing hypotheses (0.1% and 0.65% variations of the solar constant for the Maunder minimum) and confirms a large role of solar forcing on centennial scale temperature changes. With natural forcings only, the 20th century northern hemisphere temperature fluctuations remain below 0.2°C.

Part of the uncertainty associated with future climate change is due not only to climate sensitivity but also to the carbon cycle-climate feedbacks, illustrated by an additional 21st century warming of 0.5°C by the IPSL model and 2.1°C warming by the Hadley Centre models. In other words, the probability of a large climate sensitivity increases when the carbon feedbacks are included. The Bern simple climate and carbon cycle model was run for the last millennium and the CO₂ simulated in response to the climate change compared with the CO₂ records from ice cores. A 6 ppm CO₂ decrease is indeed observed during the Maunder Minimum, from 16th to 18th centuries. However, the CO₂ anomaly simulated by the Bern model is not consistent with the observations, overestimating by a factor of 2 this CO₂ drop. It seems that the low frequency temperature changes are not compatible with the ice core CO₂ record. The CO₂ to temperature change slope of the Bern model (mainly resulting from changes in soil respiration rate and ocean CO₂ solubility) was used to evaluate the temperature changes during the Maunder minimum compatible with the ice core CO₂ record. The results rule out both a large temperature change and a large solar forcing for the Little Ice Age.

Pascale Braconnot focused her presentation on the mid-Holocene, 6000 years ago. This recent warm period indeed includes large reorganisations of the hydrological cycle as documented in the Sahara and has been explored by different modelling groups and paleoclimatic reconstructions within the Paleoclimate Modelling Intercomparison Project (PMIP).

The orbital forcing induces large changes in the seasonal cycle of insolation (increased summer insolation and decreased winter insolation in the Northern Hemisphere), which raises the problem of climate sensitivity definition because of space and time shifts of major patterns with respect to the seasonal radiative forcing. In addition, the larger obliquity induces slightly higher annual mean insolation at high latitudes and decreased annual mean insolation in the tropics.

Climate changes result both from the local insolation changes and the redistribution of heat due to the dynamics of the system. In particular, the increased land-sea contrast enhances the summer monsoon intensity. The coupled ocean-atmosphere models can be compared with atmosphere only simulations (with fixed modern sea-surface conditions). Coupled models show a lowering of ocean temperature due to the winter insolation and the ocean heat capacity, and therefore an increased summer land sea contrast and more intense monsoons. Such paleoclimate simulations therefore enable to study the feedbacks associated with the coupling of different climate system components.

In order to explore the mechanisms responsible for model feedbacks and the spatial structure of climate changes (northward shift of ITCZ), the radiative budgets of the HadCM2 and IPSL models were analysed. The main differences arise from their planetary albedo change, mainly due to the cloud responses. A shortwave cloud negative feedback was due to the enhanced monsoon cloudiness, whereas a cloud positive feedback was identified in mid to high latitudes, much larger for the UKMO than the IPSL model. The net cloud radiative anomalies were very similar in the two models, and their structure similar to the spatial pattern of temperature anomalies. Again, mid-Holocene climate response to orbital forcing highlights the role of cloud feedbacks in determining the amplitude and spatial distribution of climate change. No model is however able to simulate the reconstructed northward African monsoon extent, which points to an incomplete representation of the involved processes and feedbacks.

A mechanism responsible for the ITCZ shifts was identified from atmospheric models simulations. The energetics of the atmospheric column was analysed in west Africa. The model response in terms of monsoon intensification seems associated to the partitioning of energy and specifically the ratio of latent heat to total atmospheric heat budget. An analysis of the decreased monsoon at the Last Glacial Maximum showed some analogy, suggesting that the energy partitioning and model response seem robust. The same diagnostics should also be performed on future climate simulations.

Specific analyses were dedicated to the role of the ocean and vegetation by comparing atmospheric only simulations with simulations performed with coupled models including vegetation feedbacks. In the mid-latitudes, the seasonal atmospheric response is very linear with respect to the insolation forcing. Coupled simulations reveal that the vegetation change induces a spring albedo effect, and the ocean change an autumn effect, thereby transferring a seasonal forcing into an annual mean, and inducing a non linear response to the forcing, in better agreement with biome reconstructions.

Last, the role of obliquity is explored by comparing the mid-Holocene and the glacial inception (reversed obliquity situation) simulations performed with the coupled IPSL model. In the two cases, the changes annual mean latitudinal temperature gradients are associated with a positive sea-ice/albedo/temperature feedback, with consequences on the salinity. In parallel, the equator to pole insolation gradient induces a latent heat feedback (advection of moisture by the atmosphere), which changes in the Arctic runoff and interacts with the salinity–sea-ice dependency. The last glacial inception latitudinal sea-surface-temperature is in good agreement with paleoceanographic reconstructions; new synthesis of marine data are necessary for the mid-Holocene.

The range of transient temperature response at the end of the 21st century raises the concern of abrupt responses of the climate system, including the possibility of partial Greenland melt and its consequences (large sea level rise; thermohaline circulation changes). **Jonathan Overpeck** presented the last interglacial period, about 130,000 years ago, as a key period to test the ability of climate models to simulate warmer climates, and to quantify the range of Arctic temperatures and thresholds associated with a significant Greenland melt (+3 to +6 m of sea-level rise for the last interglacial period). A motivation for this work lies in the 2.9% per decade recent decrease of Arctic sea-ice, including step function reductions during autumns 2002 and 2003, as well as an observed 16% increase in the melt area around Greenland since 1979.

Last interglacial climate simulated by the NCAR CCSM2 coupled model were compared with simulations of Arctic temperature and sea-ice changes in a future with a $3\times\text{CO}_2$. With the 130 ka BP orbital forcing ($+70\text{ W/m}^2$ in summer), Greenland is 2 to 5°C warmer than at present, and sea-ice cover is reduced by 50% in autumn. A $3\times\text{CO}_2$ climate change is associated with even larger changes in sea-ice and temperatures around Greenland. It is therefore suggested that the future CO_2 increase may enable to reach quickly the threshold of temperature required to melt the Greenland ice sheet even before year 2100. This risk is further strengthened by new ice sheet dynamical processes discovered in polar ice caps and involving rapid destabilisation from downward propagation of surface melt water.

Paleoclimatic data suggest that the CCSM2 model systematically underestimates the last interglacial Arctic warming, without however including vegetation feedbacks. A small rapid event is recorded during the penultimate deglaciation, suggesting a limited reaction of the thermohaline circulation. The discussion raised the problems of Greenland contribution to sea level change during the penultimate deglaciation, the limits of the analogy between greenhouse forcing and orbital forcing with a large seasonal aspect, and the role of atmospheric moisture transport in Greenland mass balance changes.

The global climate change from Last Glacial Maximum to the pre-industrial period also offers a potential to test the climate response to large CO_2 changes (typically from 200 to 280 ppm) and to surface property changes. **Bette Otto-Bliesner** performed a variety of sensitivity experiments with the NCAR CCSM3 model, either with a slab or coupled ocean. Discrepancies in the first coupled LGM simulations already performed with the HadCM3, CSM1, CCMA and MRI models raised the concern of having similar methodologies (ocean mixing, flux correction, runoff...), and a standardized methodology was defined within PMIP2 (Paleoclimate Modelling Intercomparison Project).

Preliminary estimates of the LGM radiative forcing suggest the following distribution: -2.3 W/m^2 due to the CO_2 decrease, -0.4 W/m^2 for the other trace gases, and about -3.5 W/m^2 due to ice and vegetation changes, therefore reaching an estimate of $6-7\text{ W/m}^2$ of total forcing, ignoring dust and aerosol changes.). The slab CCSMA3 suggests slightly different sensitivities to $2\times\text{CO}_2$ (modern to future: $+2.5^\circ\text{C}$) and to LGM to pre-industrial CO_2 change ($+2.8^\circ\text{C}$). About half of the LGM cooling (here 5.7°C) is therefore due to the CO_2 decrease. The symmetry between past and future temperature anomalies is striking, even at the ocean surface, and could be associated with low cloud feedbacks.

Although temperature changes simulated with slab and coupled models have similar global averages, they differ in their spatial structure. The coupled model simulation shows a larger cooling in the tropics and the Southern Hemisphere, in contrast with warmer conditions in the Northern Hemisphere. Paleoceanographic sea-surface-temperature reconstructions from the tropical Pacific suggest a 2 to 4°C cooling and future work should be dedicated to understanding the apparent east to west Pacific gradient. The meridional overturning of CCSM3 and CSM1 is shallower and weaker than nowadays, which is broadly consistent with proxy records from the north Atlantic, and seems to arise from the ocean vertical stability and Antarctic water formation.

These results highlight the interest of comparing exactly the same models for future climate change and for the Last Glacial Maximum, for which key paleoclimatic reconstructions are available to evaluate the realism of the spatial structure of climate change and therefore the associated sensitivity.

Some of the available climate models perform correctly to represent present day climate during the 20th century, but have very different $2\times\text{CO}_2$ sensitivities. In order to explore the benefit of Last Glacial Maximum to constrain the ranges of model climate sensitivity, **Stefan Rahmstorf** explored ensemble simulations performed with perturbed physics of the intermediate complexity CLIMBER model. A thousand different model versions were obtained by varying 11 parameters within the range of experts estimates (2 parameters for ocean mixing and 9 parameters associated with the representation of atmospheric feedbacks); some of these parameters were combined in order to explore extreme sensitivities. 3500 years of pre industrial equilibrium and 70 years of transient doubling CO_2 experiments were conducted as well as 3000 years of doubled CO_2 climate and 3500 years of LGM climate.

A wide range of sensitivities was therefore explored and only the 10% of the model versions which were consistent with modern day climate constraints were used for the LGM analyses. LGM boundary conditions included a decreased atmospheric CO_2 (180 ppm), a fixed vegetation cover, and dust forcing extracted from an ECHAM5 run. In all the cases, the glacial CO_2 decrease accounted for 30 to 40% of the simulated tropical cooling, and about 45% of Antarctic cooling.

In the CLIMBER simulations, tropical and polar Last Glacial Maximum temperature changes were systematically linearly related to each model version $2\times\text{CO}_2$ climate sensitivity. Keeping only model versions leading to a cooling of 2°C in the tropics and 8°C cooling in Antarctica constrained the range of $2\times\text{CO}_2$ climate sensitivities to 1.5 to 3.6°C. S. Rahmstorf summarized that such simulations seem to constrain the extreme ranges of sensitivity; the results remain however sensitive to the uncertainty in dust forcing. Decreasing the dust forcing by a factor of 2 resulted in a climate sensitivity range still below 5°C.

The discussion focused on the methodology to quantify the climate sensitivity range (error bars) and also to the linearity of glacial temperature with respect to $2\times\text{CO}_2$ climate sensitivity, which should be tested with coupled general circulation models.

4 Thematic Plenary Sessions

4.3 Modern Observations and Deduced Climate Sensitivity

Rapporteur: George Tselioudis

The first speaker of the session was **Cath Senior** of the UKMO Hadley Centre who gave a report on Results from the Exeter Workshop on Observations, Feedbacks and Climate Sensitivity. The workshop focused on the issue of cloud feedbacks and examined methods to derive metrics that can be used to quantify feedback strength and to evaluate model skill in simulating cloud feedback processes. The need was stressed in the meeting to produce common diagnostics from all model intercomparison and evaluation projects and to include more detailed diagnostics like satellite data simulators. The application of several mathematical techniques on model and observational data was suggested that could isolate and quantify the strength of the different components of the cloud feedback cycle. Those techniques include the application of neural networks on perturbed physics ensemble runs to relate errors in climate simulations to uncertainty in climate sensitivity, the application of compositing techniques on model and observational data to isolate the effects of different dynamical and thermodynamical processes on cloud properties, and the use of clustering techniques to characterize the properties and frequency of occurrence of the main cloud structures in models and observations. Emphasis was given to the calculation by climate modelling groups of the temporal evolution of effective climate sensitivity and of the geographical patterns of forcing and feedback parameters. The value of simulating and studying well-observed past climate events (such as the Pinatubo eruption) was stressed and it was suggested that such simulations should be a required ‘test’ for all models to pass. Finally, the danger of relying heavily on models to define climate sensitivity metrics without a good idea of model-observation discrepancies was pointed out, as well as the danger of relying on complicated statistical techniques and losing the focus on physical mechanisms.

Alan Robock of Rutgers University and Piers Forster of the University of Reading gave a presentation on the Use of observations from the 1991 Mt. Pinatubo eruption to estimate climate sensitivity. They showed results from several studies that emphasize the point that volcanic radiative forcing is not uniform in space or time and looking only at the global average response to volcanic eruptions makes it hard to distinguish the effects of mechanisms like the NAO, ENSO or the QBO. In their studies they used the global average temperature response to episodic volcanic eruptions to estimate climate sensitivity and used the response to the 1991 Pinatubo eruption to examine the water vapor/greenhouse feedback. They concluded that using the global average response to volcanic eruptions, the global climate sensitivity (DT_{2x}) can be constrained to approximately 1.5–4.5 K limits, with the mean about 2.5–3.0 K. The results preclude a sensitivity of less than 1.5 K. Their results, corroborated by those of other studies, showed that the observed water vapor response to the 1991 Pinatubo eruption confirms model simulations of a strong positive water vapor–greenhouse feedback, since such a feedback is needed to simulate the observed cooling. They cautioned, however, that the climate sensitivity derived from volcanic eruptions may not be representative of that from long-term anthropogenic climate change, since it ignores long time scale processes, such as the ocean circulation, and it is not always clear how to remove natural and other forced variability.

Richard Allan and Tony Slingo of Reading University and Mark Ringer of the UKMO Hadley Centre presented a study on exploiting observations of water vapour to investigate simulations of water vapour feedback processes. The main objective is to use present day satellite data to look at water vapour variability at different time and space scales and use the results to estimate potential global warming water vapour responses. They explored the use of reanalyses datasets and concluded that they are currently unsuitable for detection of subtle trends associated with water vapour feedbacks. They suggested the use of satellite observations to calculate the change of outgoing longwave radiation (OLR) with surface temperature as a proxy for water vapour feedback. They found that models may capture this change correctly but with different compensating water vapour and lapse rate feedbacks. They concluded that despite the fact that models capture well column water vapour sensitivity (which comes mostly from low-level water vapour changes) and agree with observations that tropical mean relative humidity varies only slightly on a decadal time-scale, they still appear to underestimate the variation of tropical mean cloudiness which leads to smaller variations in the radiative budget.

Keith Williams of the UKMO Hadley Centre presented a talk on Observational constraints on cloud feedback and climate sensitivity. The basic premise was that traditional model evaluation methods have not reduced the uncertainty in climate sensitivity and, therefore, we need to understand the key mechanisms at work and develop other evaluation metrics which can be demonstrated to be related to the cloud response to climate change. He presented results from model evaluation studies using regime compositing and clustering techniques and showed that by analysing clouds in regimes, we may be able to relate aspects of the GCM cloud response to climate change to present-day processes. If so, these aspects of the present-day simulations can be evaluated against observational data. As an example he showed that in the Hadley Centre model present day cloud changes with SST in the tropics can be used as a proxy for climate change and the model captures well present day variability. The question is, however, how general are the model-based relationships between climate change and present day processes. He pointed out that the Cloud Feedback Model intercomparison Project (CFMIP) will look at applying various compositing techniques to stratify GCM control and $2\times\text{CO}_2$ experiments. This will (hopefully) permit a more detailed understanding of the differences between models in the cloud response to climate change, provide a useful testbed for compositing techniques, and allow an assessment to be made as to what extent methods relating cloud response to climate change to present day processes, work amongst GCMs (and hence whether they might be relevant for the real world).

Alex Hall of UCLA presented a talk on the surface albedo feedback. The presentation addressed the issue of the surface contribution to planetary albedo variations. The study that was presented used ISCCP-based radiative flux data to separate planetary albedo changes into surface and cloud contributions and their residual and covariance terms and examined the relative importance of each contribution in different planetary regions. He concluded that the surface contribution to planetary albedo variability in ISCCP is significant everywhere except for the ice-free oceans. It is dominant in the SH sea ice zone year around, and in the other cryosphere regions for most of the year. If surface albedo were to vary in cryosphere regions in the future, the planet's shortwave radiation absorption would be significantly affected. He performed a similar analysis on output from current climate runs with the Community Climate System model and found that CCSM3 has substantially less surface albedo variability than ISCCP, particularly in the interior

of ice and snow packs, resulting in a much smaller contribution of the surface to planetary albedo variability. This may suggest ways to improve the model's surface albedo parameterization.

Filipe Aires of LMD and NASA/GISS and William Rossow of NASA/GISS presented a study on the estimation of sensitivities based on observations and model outputs. The objective is to develop a tool that is able to analyze sensitivities (i.e. partial derivatives) using actual observations and validate models in a better way than comparing mean-fields, climate prediction indices or metrics. The tool that is used is a neural network that has the advantage of estimating nonlinear sensitivities between multiple variables. They presented results from an analysis of multiple years of ISCCP and reanalysis data of cloud properties, radiative budget components and dynamical and thermodynamical parameters. They demonstrated the ability of the neural network technique to derive a very compact statistical representation of the datasets, to quantify the strength and provide a hierarchy of the parameter relationships, to map the interaction of variables, and to define quantitative validation metrics. They showed the advantages of using even a bivariate analysis over the use of linear relationships. They indicated that they plan to extend the analysis to use additional variables, to compare and model and observational results and evaluate model relationships, to examine the dependence of the results on the time resolution that is used, and to investigate how the relationships change in forced climate situations.

George Tselioudis of NASA/GISS and Columbia University presented a study on the Potential radiative and hydrologic climate feedbacks resulting from changes in mid-latitude storm characteristics. He presented dynamic regime based composites of cloud and radiative properties that showed large changes in those properties between weak, medium, and strong midlatitude storms. This is particularly relevant to climate sensitivity, since midlatitude storm frequencies and strengths appear to be changing in the last 50 years and are predicted to change even more in climate warming situations. He showed the results of simple calculations that indicate that decreases in midlatitude storm frequency and increases in storm strength similar to those predicted by climate models could produce net radiative responses as large as $6-7 \text{ W/m}^2$. This makes it important that climate models capture the observed sensitivities of cloud and radiation fields to changes in midlatitude storm characteristics. Therefore he proposed a comprehensive study to evaluate climate model cloud and radiative changes with dynamic regime.

Suki Manabe of Princeton University and Yoko Tsushima of the Frontier Research System for Global Change presented a study entitled radiative damping of annual temperature variation: testing simulated feedback. They suggested the use of the annual variation of the outgoing fluxes of the solar and thermal radiation at the top of the atmosphere from the ERBE experiment as a means to calculate the total feedback gain factor and use it to compare observed and simulated gain factors of annual variation. This would provide an estimate of systematic biases in model feedbacks. Then they proposed to perform a regression analysis between global warming and annual variation feedback factors obtained from the various IPCC models. On the regression line, one could find the value of global warming feedback factor that corresponds to the annual variation based upon ERBE. That will give a good estimate of the global warming feedback factor. The same analysis can be subsequently done for the gain factors of individual feedbacks like the cloud, water vapor, and albedo feedbacks. The analysis will evaluate the ability of models to simulate feedback strengths.

4 Thematic Plenary Sessions

4.4 Forcing, Feedback, and Radiation Codes

Rapporteur: Laurent Li

The 1997/98 El Niño: A test for climate models

Bob Cess

The El Niño signal is a major climate signal and should be correctly produced by climate models in order to get a good confidence for their predicted climate sensitivity. The 1997/98 El Niño event is a good test since it is a rather strong event and a well-observed one. This is of course a necessary but not sufficient condition for climate models to produce correct climate sensitivity for a radiative perturbation. The data used in this study reveal the characteristics of either cloud structure or atmospheric circulation. Cloud structure changes are from SAGE II cloud profiling (DJFMA cumulative cloud frequency profiles for individual years). Cloud radiative forcing is also averaged for JFMA from different satellite missions: 1985–1989 (ERBE), 1998 (CERES/TRMM) and 2001–2003 (CERES/Terra). Atmospheric circulation changes are revealed through NCEP/NCAR reanalysis. The HadAM3 model is also used to illustrate the approach through the comparison of two 10-member ensemble runs. The first one is run with climatological SST (obtained for the period 1961–1990). The second one is run with observed SST for the period January 1, 1997 to August 31, 1998.

From the pressure-longitude cross sections showing the zonal wind averaged for 5°S to 5°N, it is clear that a normal year is dominated by a strong Walker circulation over the equatorial Pacific and there is a collapse of this circulation for the 1997/98 El Niño event. This characteristic revealed by the reanalysis dataset can be relatively well produced by the HadAM3 model. Cloud radiative forcing is defined as the influence of clouds on the radiative flux at the top of the atmosphere and it is divided into two components: shortwave cloud radiative forcing and longwave cloud radiative forcing. The influence of cloud amount can be removed by considering the ratio $N = -(SW\ CRF)/(LW\ CRF)$. The case of $N = 1$ implies a cancellation of SW/LW. For $N > 1$, SW cooling effect dominates and the net CRF is negative. For $N < 1$, LW warming dominates and the net CRF is positive. In order to well show the characteristics of the Walker circulation collapse, it is interesting to define two regions and to show the averaged quantities for the two regions. The two contrast regions are naturally defined over the Eastern and Western basins of the equatorial Pacific.

For the two selected regions, the SAGE II cloud profiling curves showing the cumulative cloud frequency of DJFMA are plotted for normal years (averaged from 1985 to 1991) and the El Niño year 1998. The decrease of cumulative cloud frequency for the western region can be about 40% and the increase for the eastern region is even larger (55%). The ratio N is also plotted year-by-year from 1985 to 2003 for the two regions respectively. For normal years, the ratio N for the western Pacific is around 1.1, showing a near cancellation between the SW and LW CRF; the ratio N is between 1.5 and 2.1 for the eastern region, showing a strong cooling effect. It is clear that the ratio N is in general smaller over the Western Pacific than over the Eastern Pacific. But

for the year 1998, the reverse situation is obtained, with the value of N slightly stronger in the Western Pacific (1.5) than in the Eastern Pacific (1.4). The comparison with HadAM3 reveals a very good model performance in reproducing CRF characteristics for both the Western and Eastern regions. It is suggested to generalize this comparison to other models as a constrain condition for their reliability in climate sensitivity studies.

Geographic aspects of climate feedback and sensitivity

George Boer

Currently, climate sensitivity is considered as a single parameter showing the ratio of global-averaged surface air temperature change over the globally-averaged radiative perturbation. It is suggested that a locally-defined climate sensitivity parameter $L(x,y,t)$ is a more appropriate concept to characterize the relationship of forcing $F(x,y,t)$ and response $T(x,y,t)$. Different forcing patterns with the same global mean $\langle F \rangle$ may give different global mean temperature responses $\langle T \rangle$. This is because forcing preferentially in regions of positive feedback gives a larger global mean temperature response than does forcing uniformly or forcing preferentially in regions of negative feedback. This “efficacy” of the forcing arises from the local nature of the feedbacks as measured by $L(x,y,t)$. From an experiment of IS92a scenario (greenhouse gas and aerosol concentrations stabilized at year 2050) with the model CCCma CGCM2, the local feedback parameter $L(x,y,t)$ is calculated as demonstration. The high latitude regions of positive feedback are associated with the surface ice/snow albedo feedback while those in the tropical Pacific depend on changes in the structure of moisture, temperature and cloudiness. Negative regions are associated with increased outgoing IR from the warmer surface and atmosphere and with decreased solar flux associated with changes in cloud distribution and optical properties.

Feedback parameter $L(x,y,t)$ can be decomposed into components such as $L = L_s + L_l = L_a + L_c = L_{sa} + L_{sc} + L_{la} + L_{lc}$, associated respectively with solar (s), longwave (l), clear-sky atmosphere/surface (a) and cloud (c) feedbacks. This offers a powerful tools for intercomparing models and for understanding their responses, as illustrated by three models involved (CSM1.4, CSM2.0 and CGCM3).

Climate sensitivity for different forcings: Results from METRIC project

Robert Sausen

The main objective of the METRIC project is to understand climate sensitivity in response to geographically inhomogeneous forcings, such as aerosols and short-lived greenhouse gases. The experiment protocol is quite simple and realized by atmospheric GCMs coupled to a slab ocean. A variety of geographical patterns of the forcing are explored with idealized distributions over the globe: tropics, extratropics and different combinations. Forcings with different nature are also explored: CO_2 , solar radiation, CH_4 and O_3 . For all experiments, forcings are scaled to 1 W/m^2 for global average. It seems that the surface albedo feedback is the dominating one and the stratospheric water vapour feedback in the case of O_3 is a very strong positive feedback,

although the troposphere/stratosphere exchange is still a quite uncertain issue. Tropospheric O_3 forcings due to emission of precursors, such as NO_x and CO , in different regions of the world are also studied. Finally, the METRIC experimental protocol can be extended to a larger number of modelling groups and this will provide a suitable framework to examine feedbacks in climate models.

Solar and longwave radiative transfer: LBL, GCMs and observation

V. Ramaswamy

A comparison is done among different models (LBL, NBM, WBM) for short- and longwave radiative transfer. For the clear-sky longwave transfer, trace gases and water vapour continuum are two important issues. LBL is in general considered as the reference and for a MLS (Mid-Latitude Summer) atmosphere, an increase of CO_2 (from 287 to 574 ppm) gives a radiative forcing of 5.48 W/m^2 at the tropopause, an increase of CH_4 (from 0 to 1.76 ppb) gives 4.60 W/m^2 and an increase of N_2O (from 0 to 0.316 ppb) gives 4.42 W/m^2 . For most NBM and WBM, the accuracy is within 5% in comparison to LBL in the case of CO_2 doubling. For shortwave transfer, intercomparison works show that current models give a range of several percents for the accuracy of calculations in the cases of molecular atmosphere, aerosol skies and cloud-filled skies. Effect of water vapour continuum and trace gases on the shortwave transfer can be also important. Estimation of SW forcing due to increase of CO_2 and CH_4 from 1860 to 2000 is respectively 0.27 and 0.13 W/m^2 for the tropopause. The following conclusions can be drawn: (1) Improved radiation parameterizations for trace gases and water vapor continuum will bring improvements in climate simulations. (2) Intercomparisons, with the help of LBL computations, adding a degree of confidence in the evaluation of codes. (3) Several types of observations available to cross-check SW and LW at least, clear-sky could be made more robust. (4) Methane solar absorption as significant, if not more than solar CO_2 .

State of radiation codes in GCMs

Qiang Fu

A survey of the radiation codes used in current GCMs has been done and the main purpose is to document their differences and to narrow the uncertainty of climate prediction. An official report of this survey will be soon released to the climate community. Water vapour continuum absorption, light scattering by non-spherical ice particles, inhomogeneity of clouds and cloud vertical overlap are the main issues of the current radiation codes.

From this survey, the following points seem to be the major challenges in radiation parameterizations for GCMs: Treatment of cloud inhomogeneities, real coupling of cloud and radiation, improving accuracy and reducing computing time, lateral inhomogeneity in cloud distributions and condensed water, ice cloud optical properties and cloud overlapping, and inclusion of sub-grid scale clouds and realistic treatment of cloud overlap.

Also from the survey, the following points seem the priorities for further developments:

- Having new treatment of non-spherical ice particles; including continuum in SW; using predicted particle size from prognostic aerosol concentrations; revising absorption by using H2000, CKD2.4, and SW newly discovered weak lines
- Including non-LTE effects for O₃ and CO₂
- Improvement of overall precision using updated absorption coefficients; incorporation of radiatively interactive aerosols.
- Including most recent CKD; scattering in LW; inclusion of longwave effects of stratospheric sulfuric acid, dust.
- Improving ice cloud optical properties and cloud overlapping. Taking cloud overlap assumption out of radiation scheme and using the new cloud overlap in terms of correlation length between cloud layers which is consistent for both radiation and precipitation/evaporation processes.
- Move to use modified Edward-Slingo code. More spectral bands with better absorption models.
- Including ice cloud optical properties and improving cloud overlap treatment.
- Inclusion of NO₂, O₂*O₂, and O₂*N₂ in SW; inclusion of non-LTE for CO₂; interactive determination of aerosol properties.

Comparison of radiative forcings from GCM and LBL models

William Collins

The goals of this radiation codes intercomparison project are (1) to compare forcing by well-mixed GHGs from GCMs participating in the IPCC AR4, and (2) to determine the accuracy of GCM codes under idealized conditions. The forcing considered is the difference between present-day and the pre-industrial, doubling of CO₂, and feedbacks from increased H₂O. Concerning the design of the comparison experiments, a typical mid-latitude Summer profile (T , q , O₃ and P) is used with 40 levels for GCMs and 458 levels for the LBL models. For well-mixed GHGs, volumic uniform dry mixing ratio is prescribed. As boundary conditions, the surface emissivity is set to 1, lambertian surface albedo is set to 0.1, the solar constant is set to 1360 W/m² and the zenith angle is 53°. The comparison is done for instantaneous forcing without stratospheric adjustment. This is because instantaneous forcings are included in WGCM protocol for IPCC simulations. Calculations are for clear-sky conditions only, since inclusion of clouds would complicate too much the comparison. The following effects are included: absorption by H₂O, O₃, and well-mixed GHGs; Rayleigh scattering; self and foreign line broadening. Eight experiments are currently proposed. Five GCM groups and six LBL model groups are

now involved in the comparison. Other groups are welcome to join this project. Preliminary results show that there is no error in sign. Most GCMs don't include the shortwave effect of CH₄. Following spread in flux estimates are obtained: 3.45 ± 0.98 W/m² (longwave) for the top of the model in the case of CH₄ + N₂O; 1.08 ± 0.71 W/m² (longwave) and 0.85 ± 0.76 W/m² (shortwave) for the surface in the case of CO₂. More participating GCM groups are necessary to make the statistics robust.

5 References

- Andronova, N.G. and M.E. Schlesinger, 2001: Objective estimation of the probability density function for climate sensitivity. *J. Geophys. Res.*, **106**, 22605–22611.
- Boer, G.J., and B. Yu, 2003: Climate sensitivity and climate response. *Climate Dynamics*, **20**, 415–429.
- Bony, S., J.-L. Dufresne, H. Le Treut, J.-J. Morcrett, and C. Senior, 2004: On dynamic and thermodynamic components of cloud changes. *Climate Dynamics*, **22**, 71–86.
- EPICA community members, 2004: Eight glacial cycles from an Antarctic ice core. *Nature*, **429**, 623–628.
- Forest, C.E., P.H. Stone, A.P. Sokolov, M.R. Allen and M.D. Webster, 2002: Quantifying uncertainties in climate system properties with the use of recent climate observations. *Science*, **295**, 113–117.
- Forster, P. M. de F. and M. Collins, 2004: Quantifying the water vapour feedback associated with post-Pinatubo global cooling. *Climate Dynamics*, DOI: 10.1007/s00382-004-0431-z.
- Frame, D.J and M.R. Allen, 2004: Observational constraints and prior assumptions on climate sensitivity. *Extended abstract in this volume*.
- Gregory, J.M., R.J. Stouffer, S.C.B. Raper, P.A. Stott and N.A. Rayner, 2002: An observationally based estimate of the climate sensitivity. *J. Climate*, **15**, 3177–3121.
- Hansen, J., A. Lacis, R. Ruedy, M. Sato, and H. Wilson, 1993: How sensitive is the world's climate? *Natl. Geog. Res. Exploration*, **9**, 142–158.
- Hegerl, G.C. and T.J. Crowley, 2004: Estimating temperature response to radiative forcing from paleoclimate records of the last millennium. *Extended abstract in this volume*.
- Hoffert, M. I., and C. Covey, 1992: Deriving global climate sensitivity from paleoclimate reconstructions. *Nature*, **360**, 573–576.
- Knutti, R., T.F. Stocker, F. Joos and G.K. Plattner, 2002: Constraints on radiative forcing and future climate change from observations and climate model ensembles. *Nature*, **416**, 719–723.
- Lorius, C., J. Jouzel, D. Raynaud, J. Hansen, and H. Le Treut, 1990. The ice-core record: Climate sensitivity and future greenhouse warming. *Nature*, **347**, 139–145.
- Moss, R. and S. Schneider, 2000: Uncertainties. In Guidance Papers on the Cross Cutting Issues of the Third Assessment Report of the IPCC. R. Pachauri, T. Taniguchi and K. Tanaka, eds. IPCC, Geneva.
- Murphy, J., 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. *Journal of Climate*, **8**, 496–514.
- Murphy, J.M, D.M.H. Sexton, D.N. Barnett, G.S. Jones, M.J. Webb, M. Collins, and D.A. Stainforth, 2004: Quantification of modeling uncertainties in a large ensemble of climate change simulations. *Nature*, **430**, 768–772.
- Rahmstorf, S., D. Archer, D.S. Ebel, O. Eugster, J. Jouzel, D. Maraun, U. Neu, G.A. Schmidt, J. Severinghaus, A.J. Weaver, and J. Zachos, 2004: Cosmic rays, carbon dioxide and climate. *EOS Transactions of the AGU*, **85**(4), 38, 41
- Räisänen, J., 2004: Probability distributions of CO₂-induced global warming as inferred directly from multi-model ensemble simulations. *Extended Abstract in this volume*.
- Senior, C., B. Wielicki, B. McAvaney and G. Boer, 2004: Report on the joint WCRP CFMIP/

- IPCC expert meeting on climate sensitivity and feedbacks. *See Annex 5 in this volume.*
- Shaviv, N.J. and J. Veizer, 2003: Celestial driver of Phanerozoic climate? *GSA Today*, **13**(7), 4–10.
- Shine, K. P., B. P. Briegleb, A. S. Grossman, D. Hauglustaine, H. Mao, V. Ramaswamy, M. D. Schwarzkopf, R. Van Dorland, and W-C. Wang, 1995: Radiative forcing due to changes in ozone: A comparison of different codes. In *Atmospheric Ozone as a Climate Gas*, NATO ASI Series I, Vol. 32, Heidelberg, Germany: Springer-Verlag, 373–396.
- Shine, K.P., J. Cook, E.J. Highwood, M. Joshi, 2003: An alternative to radiative forcing for estimating the relative importance of climate change mechanisms. *Geophys. Res. Lett.*, **30**(20), doi:10.1029/2003GL018141.
- Soden, J.B., R.T. Wetherald, G.L. Stenchikov, and A. Robock, 2002: Global cooling after the eruption of Mount Pinatubo: A test of climate feedback by water vapor. *Science*, **296**, 727–730.
- Stainforth, D.A. et al., 2004: Evaluating Uncertainty in the Climate Response to Changing Levels of Greenhouse Gases. *Submitted.*
- Stott, P.A. and J.A. Kettleborough, 2002: Origins and estimates of uncertainty in predictions of twenty-first century temperature rise. *Nature*, **416**, 723–726.
- Tebaldi, C., R.L. Smith, D. Nychka and L.O. Mearns, 2004: Quantifying uncertainty in projections of regional climate change: A Bayesian approach to the analysis of multi-model ensembles. *Submitted.*
- Von Deimling, T.S., H. Held, A. Ganopolski and S. Rahmstorf, 2004: Climate sensitivity range derived from large ensemble simulations of glacial climate constrained by proxy data. *Extended abstract in this volume.*
- Wigley, T.M.L. and S.C.B. Raper, 2001. Interpretation of high projections for global-mean warming. *Science*, **293**, 451–454.

Annex 1:

Workshop Program

DAY 1: Monday, 26 July

08:00–08:30 REGISTRATION

Introduction and Overview Session

(Chairs: J. Jouzel and S. Bony; Rapporteur: C. Ammann)

08:30–09:00	<i>Sénateur Paul Vergès</i>	Welcome on behalf of Observatoire National sur les Effets du Réchauffement Climatique (ONERC)
	<i>Sylvie Joussaume</i>	Welcome on behalf of Institut National des Sciences de l'Univers at CNRS (INSU)
	<i>Michael Ghil</i>	Welcome on behalf of École Normale Supérieure (ENS)
	<i>S. Solomon and SSC Co-chairs</i>	Introduction and welcome, purpose of the workshop, logistics.
09:00–09:30	<i>J. Gregory</i>	Review of concepts and methods relating to climate sensitivity
09:30–10:00	<i>R. Colman</i>	Understanding and comparing climate feedbacks in models
10:00–10:30	<i>S. Raper</i>	Interpretation of model results that show changes in the effective climate sensitivity with time
10:30–11:00	BREAK	
11:00–11:30	<i>J. Murphy</i>	Probabilistic estimates of climate sensitivity from GCM ensembles
11:30–12:00	<i>J. Jouzel</i>	Climate sensitivity inferred from paleoclimatic records
12:00–12:30	<i>S. Bony</i>	How can we constrain the cloud-radiative feedback produced by climate models?
12:30–14:00	LUNCH	

Climate Sensitivity from Models

(Chairs: G. Meehl, B. McAvaney; Rapporteur: R. Allan)

14:00–14:30	<i>J. Kiehl</i>	A comparison of climate sensitivity between the NCAR and GFDL models
14:30–15:00	<i>H. Le Treut and B. McAvaney</i>	Issues about climate sensitivity: Results from LMD and BMRC
15:00–15:15	<i>R. Stouffer</i>	Climate sensitivity: Atmosphere-mixed layer ocean model vs AOGCM
15:15–15:30	<i>A. Broccoli</i>	Diagnosing climate sensitivity and feedbacks in the GFDL AM2 model
15:30–16:00	BREAK	
16:00–16:30	<i>G. Meehl</i>	Ocean heat uptake and climate sensitivity in five coupled models
16:30–17:00	<i>A. Kitoh</i>	A comparison of climate sensitivity among the Japanese models

Reception

Hosted by CNRS

DAY 2: Tuesday, 27 July**Forcing, Feedback, and Radiation Codes***(Chairs: W. Collins, V. Ramaswamy; Rapporteur: L. Li)*

08:55–09:20	R. Cess	The 1997/98 El Niño: A test for climate models
09:20–09:45	G. Boer	Geographic aspects of climate feedback/sensitivity
09:45–10:10	R. Sausen	METRIC: A comparison of climate sensitivity from different forcings
10:10–10:35	V. Ramaswamy	Solar and longwave radiative transfer: LBL, GCMs, and observations
10:35–11:00	BREAK	
11:00–11:25	Q. Fu	State of radiation codes in GCMs: Some basic aspects
11:25–11:50	W. Collins	Comparison of radiative forcings from GCMs and line-by-line models
11:50–12:15	S. Kinne	Climate sensitivity and the role of aerosol effects
11:15–13:30	LUNCH	

Climate Sensitivity PDFs and Measures*(Chairs: C. Senior, G. Boer; Rapporteur: M. Collins)*

13:30–13:55	N. Andronova	Aerosol forcing and PDFs of climate sensitivity
13:55–14:20	M. Allen	Observational constraints and prior assumptions on climate sensitivity
14:20–14:45	J. Raisanen	Probability distributions of CO ₂ -induced global warming as inferred directly from multi-model ensemble simulations
14:45–15:10	R. Knutti	Twentieth century climate and probabilistic estimates of climate sensitivity
15:10–15:35	G. Hegerl	Estimating climate sensitivity from paleoclimatic records of the last millennium
15:35–16:05	BREAK	

Paleoclimatic Observations and Climate Sensitivity*(Chairs: C. Senior, G. Boer; Rapporteur: V. Masson-Delmotte)*

16:05–16:30	F. Joos	What can be learned from variations in atmospheric CO ₂ and temperature over the past millennium?
16:30–16:55	J. Overpeck	Paleoclimatic evidence for future Greenland ice sheet instability and rapid sea level rise
16:55–17:20	B. Otto-Bliesner	Climate sensitivity of the Last Glacial Maximum from paleoclimate simulations and observations
17:20–17:45	P. Braconnot	What can we learn on climate sensitivity from mid-Holocene type simulations?
17:45–18:10	S. Rahmstorf	Climate sensitivity range derived from large ensemble simulations of glacial climate constrained by proxy data

Dinner Aboard the Ship Onyx on the Seine*Hosted by the Ministry of Ecology and Sustainable Development*

19:30–23:00 Meet at Port de la Bourdonnais, 75007 Paris, at 19:30

DAY 3: Wednesday, 28 July**Modern Observations and Deduced Climate Sensitivity***(Chairs: S. Bony, J. Gregory; Rapporteur: G. Tselioudis)*

08:30–09:00	<i>C. Senior</i>	Results from the Exeter workshop: Observations, feedbacks and climate sensitivity
09:00–09:30	<i>A. Robock and P. Forster</i>	Use of observations from the Mt Pinatubo eruption to estimate climate sensitivity
09:30–10:00	<i>R. Allan</i>	Exploiting observations of water vapour to investigate simulations of water vapour feedback processes
10:00–10:30	<i>K. Williams</i>	Observational constraints on cloud feedback and climate sensitivity
10:30–11:00	BREAK	
11:00–11:30	<i>A. Hall</i>	Controls on the interannual variability of planetary albedo and relevance for climate sensitivity
11:30–11:45	<i>F. Aires</i>	Estimation of sensitivities based on observations & model outputs
11:45–12:00	<i>G. Tselioudis</i>	Potential radiative and hydrologic climate feedbacks resulting from changes in mid-latitude storm characteristics
12:00–12:30	<i>S. Manabe</i>	Radiative damping of annual temperature variation: A proposal for testing feedback
12:30–12:45	<i>S. Solomon and SSC Co-chairs</i>	Discussion of breakout group goals, approaches and process
12:45–14:15	LUNCH	

Breakout Session 1

14:15–15:30	Breakout Groups 1 and 2 in parallel
	BOG 1: Climate sensitivity from models (Leaders: G. Meehl, B. McAvaney) Short presentations on equilibrium sensitivity and TCR from each modelling group, and how these relate to IPCC TAR numbers; what are the new ranges for equilibrium and TCR and why have these ranges changed.
	BOG 2: Climate sensitivity from observations (Leaders: S. Bony, J. Gregory) Specify the best estimate and range for climate sensitivity from observations and recommend a preliminary database of observations.
15:30–16:00	BREAK
16:00–18:00	Breakout Groups 1 and 2 continued

DAY 4: Thursday, 29 July

Breakout Session 2

- 08:30–10:00 Breakout Groups 3 and 4 in parallel
 BOG 3: Radiative transfer and forcing (*Leaders: W. Collins, V. Ramaswamy*)
 Compare and assess radiation codes and forcings used in GCMs, including globally averaged and geographic patterns of forcings.
 BOG 4: Probabilistic approaches to climate sensitivity (*Leaders: C. Senior, G. Boer*)
 Assess techniques to formulate probabilistic estimates of climate sensitivity and propose an approach to climate sensitivity evaluation (including pdfs) for the AR4.
- 10:00–10:30 **BREAK**
- 10:30–11:00 Breakout Groups finalize presentations

Closing Plenary*(Chairs: SSC Co-chairs)*

- 11:00–11:30 *B. Otto-Bliesner and J. Gregory* BOG2 Report to Plenary (20 min + 10 min discussion)
- 11:30–12:00 *G. Meehl* BOG1 Report to Plenary (20 min + 10 min discussion)
- 12:00–12:30 *W. Collins* BOG3 Report to Plenary (20 min + 10 min discussion)
- 12:30–14:00 **LUNCH**
- 14:00–14:30 *C. Senior* BOG4 Report to Plenary (20 min + 10 min discussion)
- 14:30–15:00 *R. Somerville* Key conclusions: Clouds and Radiative Transfer
- 15:00–15:30 *T. Stocker* Key conclusions: Observations and Paleoclimate
- 15:30–16:00 *D. Randall* Key conclusions: A plea from Chapter 8
- 16:00–16:30 *SSC Co-chairs and S. Solomon* Closing Remarks
- 16:30 ADJOURN**
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Water Vapour Feedback Observations and Climate Sensitivity

Richard P. Allan

Environmental Systems Science Centre (ESSC), University of Reading

The response of atmospheric moisture to changes in surface temperature (T_s) determines to a large extent the sensitivity of the climate system to a radiative perturbation. Aside from the indirect influence of moisture changes on cloud feedbacks, a primary component of the direct water vapour feedback (β_{wv}) is encapsulated by,

$$\beta_{wv} \approx \left(\frac{\partial OLR_c}{\partial WV} \right) \left(\frac{\partial WV}{\partial T_s} \right), \quad (1)$$

where OLR_c is the clear-sky outgoing longwave radiation and WV is a generic water vapour variable. An important step in diagnosing water vapour feedback from observations is therefore to establish a relationship between water vapour concentrations and the surface temperature. Although it is only possible to measure dWV/dT_s rather than $\partial WV/\partial T_s$, it is possible to reduce this difference by removing the effects of the large scale circulation on the local changes in WV . This may be achieved by subsampling dynamical regimes (e.g. Bony *et al.* (1997), Allan *et al.* (2002c)) or by averaging over the large-scale circulation systems (e.g. Allan *et al.* (2002a)).

An important theoretical constraint on the water vapour feedback is the Clausius Clapeyron equation which predicts an approximately exponential increase in water vapour with temperature where relative humidity (RH) is conserved (e.g. Raval and Ramanathan (1989)). Wentz and Schabel (2000) demonstrated an observed increase in column integrated water vapour (CWV) with T_s of about 9% K^{-1} , close to that predicted by the Clausius Clapeyron equation, by analysing trends over the ocean. In Fig. 1a-b both models and satellite observations show excellent agreement in the relationship between CWV and T_s over a decadal time-scale (see also Soden (2000)) with $dCWV/dT_s = 3.5 \text{ kg m}^{-2} (\approx 9\% K^{-1})$.

Given the strong coupling between ocean surface temperature and boundary layer water vapour, which is the primary determinant of CWV , it would be surprising if the relationship between marine CWV and T_s did not hold. However, OLR_c is sensitive to humidity changes throughout the troposphere (e.g. Allan *et al.* (1999)) so it is therefore important also to evaluate the free tropospheric moisture changes simulated by models. One possibility is to use reanalyses which assimilate a variety of observations into an atmospheric model

and output variables such as the vertical profiles of atmospheric water vapour globally. However, the changing quality of the observational input to reanalyses render the presently available products unsuitable for the analysis of water vapour feedback (Trenberth *et al.* (2001), Allan *et al.* (2002b), Allan *et al.* (2004)).

Because OLR_c is highly sensitive to humidity throughout the troposphere it is feasible to use $dOLR_c/dT_s$ as a proxy for β_{wv} (e.g. Raval and Ramanathan (1989), Slingo *et al.* (2000)). Cess *et al.* (1990) demonstrated good agreement between model $dOLR_c/dT_s$ and interpreted this as consistency in water vapour feedback. Agreement between observed and simulated variations in OLR_c (Soden (2000), Allan and Slingo (2002)) suggest that the simulated water vapour feedback is realistic. For example, Fig. 1c shows reasonable agreement between observed and model simulated normalised greenhouse trapping, $g_a = 1 - (\sigma T_s^4 / OLR_c)$, with increased greenhouse trapping during warm events, symptomatic of positive water vapour feedback (Allan *et al.* (2003)). However, as demonstrated in Fig. 1c (dashed line), g_a is also sensitive to forcings such as greenhouse gas concentration changes and volcanic aerosols which may confuse the diagnosis of water vapour feedback from analysing broadband radiative fluxes. In addition to this limitation, similarity in $dOLR_c/dT_s$ does not necessarily indicate consistency in water vapour feedback. For example, Allan *et al.* (2002a) showed that 2 models with identical forcings produced a similar sensitivity, $dOLR_c/dT_s \approx 2 \text{ W m}^{-2} K^{-1}$, but contained rather different temperature and water vapour profile responses to T_s over an interannual time-scale. The discrepancy, which was ascribed to differences in the model convection parametrizations, raises questions as how best to diagnose water vapour feedback (see also Held and Soden (2000)) and how the water vapour, temperature lapse rate and cloud feedbacks may interact.

Colman (2003) compared climate feedbacks from a variety of models and found a large compensation between water vapour and temperature lapse rate feedback, consistent with the analysis of Allan *et al.* (2002a). Based on the apparent robust nature of modelled and observed constant relative humidity water vapour feedback feedback (e.g. Ingram (2002), Soden *et al.* (2002)) it seems reasonable to check for departure from this theoretical relationship by

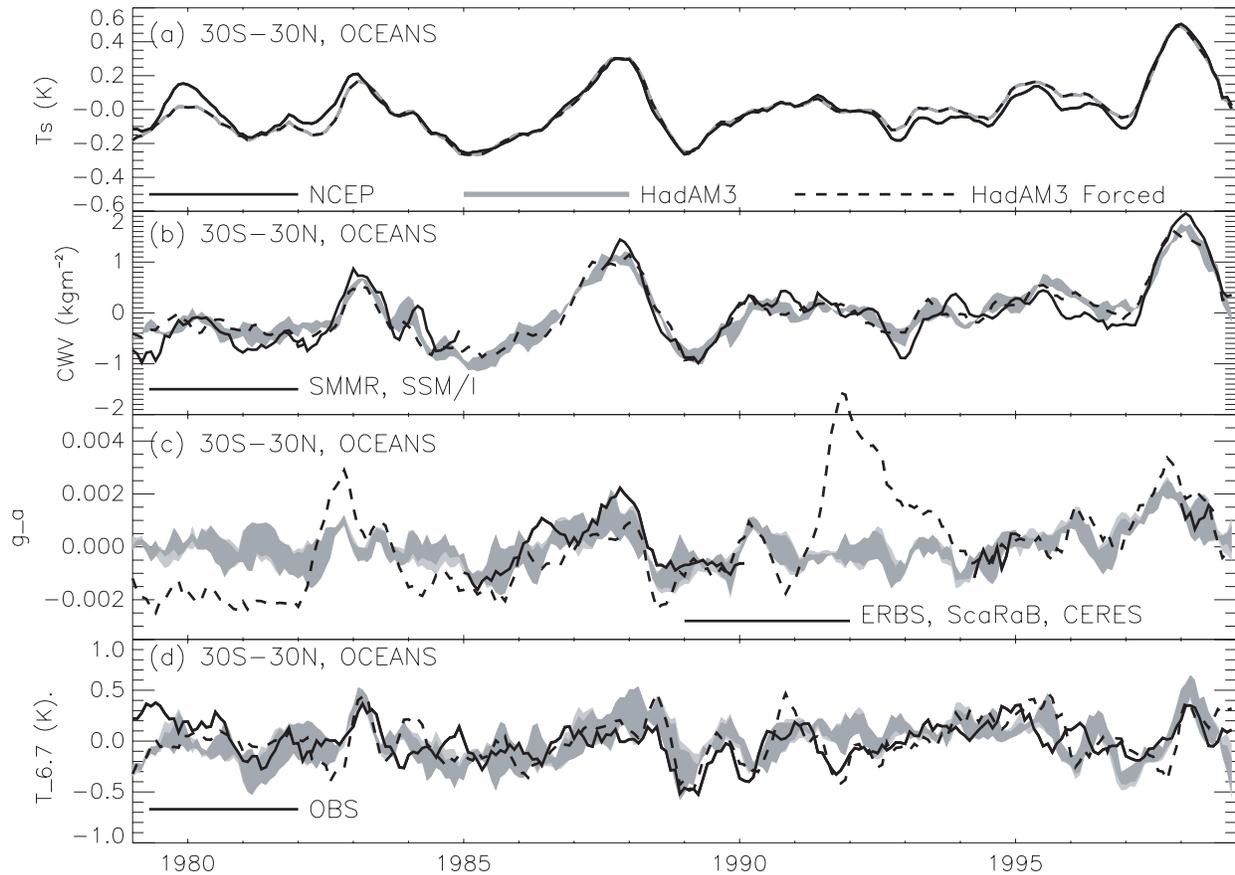


Figure 1. Interannual variations in (a) surface temperature, (b) column integrated water vapour, (c) atmospheric normalised greenhouse trapping and (d) 6.7 μm brightness temperature for sea surface temperature (SST) forced model (shaded), model with all known forcings (dashed) and observations (solid) (from Allan *et al.* (2003)).

measuring the feedback, if any, involving relative humidity. An additional benefit of this approach is the potential applicability to cloud feedbacks given the strong relationship between RH and cloudiness (e.g. J. M. Slingo (1980)). Thus it is important to evaluate the sensitivity of OLR_c to RH ($\partial OLR_c/\partial RH$) and to diagnose the changes in RH in response to T_s .

Figure 2 illustrates a technique to estimate $\partial OLR_c/\partial RH$ by computing $dOLR_c/dUTH$ using the results of Allan *et al.* (2003). Here, $dOLR_c/dUTH$ is calculated at each tropical grid-point from interannual monthly anomalies, plotted as a function of mean UTH where UTH is estimated from observations and simulations of 6.7 μm radiances. The increasingly negative $dOLR_c/dUTH$ with decreasing humidity is consistent with previous studies (e.g. Spencer and Braswell (1997)) although the model appears to overestimate the magnitude of this sensitivity, especially at low humidities compared with the combined ERBS and HIRS satellite observations. Regardless of the approximate relationship, $dOLR_c/dUTH \approx 0.5\% \text{ K}^{-1}$, the departure from a constant relative humidity water vapour feedback appears small on the interannual

time-scale because changes in 6.7 μm radiance (or equivalent brightness temperature, $T_{6.7}$) are small and not significantly correlated with T_s (Fig. 1d; Allan *et al.* (2003)). Although $T_{6.7}$ does not appear to be directly influenced by additional forcings (see dashed line in Fig. 1d) the relationship between $T_{6.7}$ and UTH may not be robust on interannual time-scales where temperature changes may also influence changes in $T_{6.7}$ (Allan *et al.* 2003). Therefore, these techniques may need to be further refined. Finally, understanding the links between T_s , RH , cloudiness and the large-scale dynamics may improve our understanding of climate feedbacks and how they interact with one another (e.g. Hartmann *et al.* (2001)).

References

- Allan, R. P. and Slingo, A. (2002). Can current climate model forcings explain the spatial and temporal signatures of decadal OLR variations? *Geophys. Res. Lett.*, **29**, 0.1029/2001GL014620.
- Allan, R. P., Shine, K. P., Slingo, A., and Pamment, J. A. (1999). The dependence of clear-sky outgoing longwave radiation on surface temperature and

- relative humidity. *Quart. J. Roy. Meteorol. Soc.*, **125**, 2103–2126.
- Allan, R. P., Ramaswamy, V., and Slingo, A. (2002a). A diagnostic analysis of atmospheric moisture and clear-sky radiative feedback in the Hadley Centre and Geophysical Fluid Dynamics Laboratory (GFDL) climate models. *J. Geophys. Res.*, pages 4329, doi 10.1029/2001JD001131.
- Allan, R. P., Slingo, A., and Ramaswamy, V. (2002b). Analysis of moisture variability in the European Centre for Medium-Range Weather Forecasts 15-year reanalysis over the tropical oceans. *J. Geophys. Res.*, **107**, 4230, doi 10.1029/2001JD001132.
- Allan, R. P., Slingo, A., and Ringer, M. A. (2002c). Influence of dynamics on the changes in tropical cloud radiative forcing during the 1998 El Niño. *J. Climate*, **15**, 1979–1986.
- Allan, R. P., Ringer, M. A., and Slingo, A. (2003). Evaluation of moisture in the Hadley Centre climate model using simulations of HIRS water-vapour channel radiances. *Quart. J. Roy. Meteorol. Soc.*, **129**, 3371–3389.
- Allan, R. P., Ringer, M. A., Pamment, J. A., and Slingo, A. (2004). Simulation of the Earth's radiation budget by the European Centre for Medium Range Weather Forecasts 40-year reanalysis (ERA40). *J. Geophys. Res.*, **accepted**.
- Bony, S., Lau, K., and Sud, Y. C. (1997). Sea surface temperature and large-scale circulation influences on tropical greenhouse effect and cloud radiative forcing. *J. Climate*, **10**, 2055–2077.
- Cess, R. D., and coauthors (1990). Intercomparison and interpretation of climate feedback processes in 19 atmospheric general circulation models. *J. Geophys. Res.*, **95**, 16601–16615.
- Colman, R. (2003). A comparison of climate feedbacks in general circulation models. *Climate Dynamics*, **20**, 865–873.
- Hartmann, D. L., Moy, L. A., and Fu, Q. (2001). Tropical convection and the energy balance at the top of the atmosphere. *J. Climate*, **14**, 4495–4511.
- Held, I. and Soden, B. J. (2000). Water vapor feedback and global warming. *Ann. Rev. Energy Environ.*, **25**, 441–475.
- Ingram, W. J. (2002). On the robustness of the water vapor feedback: GCM vertical resolution and formulation. *J. Climate*, **15**, 917–921.
- Raval, A. and Ramanathan, V. (1989). Observational determination of the greenhouse effect. *Nature*, **342**, 758–761.
- Slingo, A., Pamment, J. A., Allan, R. P., and Wilson, P. (2000). Water vapour feedbacks in the ECMWF Re-Analyses and Hadley Centre climate model. *J. Climate*, **13**, 3080–3098.

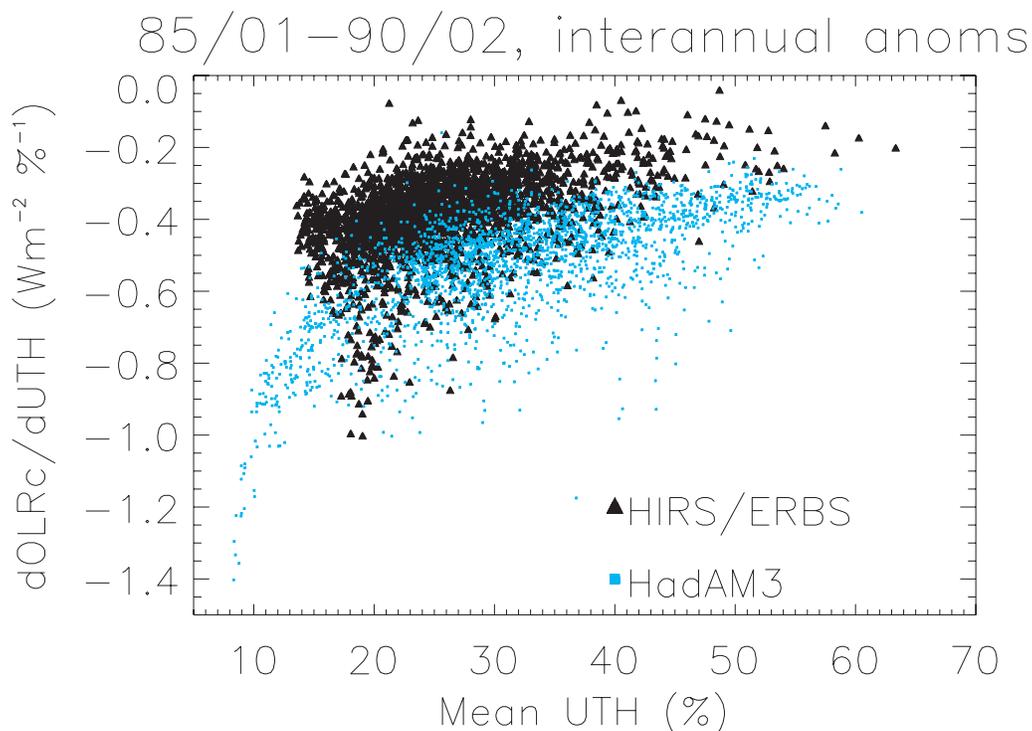


Figure 2. Sensitivity of clear-sky OLR to upper tropospheric humidity (UTH) as a function of mean UTH for the HadAM3 model and satellite observations.

- Slingo, J. M. (1980). A cloud parameterization scheme derived from GATE data for use with a numerical model. *Quart. J. Roy. Meteorol. Soc.*, **106**, 747–770.
- Soden, B. J. (2000). The sensitivity of the tropical hydrological cycle to ENSO. *J. Climate*, **13**, 538–549.
- Soden, B. J., Wetherald, R. T., Stenchikov, G. L., and Robock, A. (2002). Global cooling after the eruption of Mount Pinatubo: a test of climate feedback by water vapor. *Science*, **296**, 727–730.
- Spencer, R. W. and Braswell, W. D. (1997). How dry is the tropical free troposphere? Implications for global warming theory. *Bull. Amer. Met. Soc.*, **78**, 1097–1106.
- Trenberth, K. E., Stepaniak, D. P., Hurrell, J. W., and Fiorino, M. (2001). Quality of reanalyses in the tropics. *J. Climate*, **14**, 1499–1510.
- Wentz, F. J. and Schabel, M. (2000). Precise climate monitoring using complementary satellite data sets. *Nature*, **403**, 414–416.

Annex 2: Extended Abstracts

Aerosol Forcing and pdfs of Climate Sensitivity

Natalia G. Andronova and Michael E. Schlesinger

Climate Research group, Department of Atmospheric Sciences, University of Illinois at Urbana-Champaign, natasha@atmos.uiuc.edu and schlesin@atmos.uiuc.edu

According to recent findings, the radiative forcing by anthropogenic tropospheric aerosol may be an offset for the strongly positive radiative forcing due to anthropogenic greenhouse gases. This offset would result in higher climate sensitivity, as required to reproduce the recent trend in the observed near surface temperature. Uncertainties in the estimation of the magnitude of the aerosol forcing are greater than for the well-mixed greenhouse gases. This is mostly due to the different nature of aerosols and their ability to modify the radiative properties of other important climate agents such as clouds and ice and snow. Here we discuss the uncertainties of the aerosol forcing and how they influence the estimation of the sensitivity of the climate system.

Introduction

The 2001 IPCC Report estimated the total aerosol forcing as having a mean value of -0.5 W/m^2 and an uncertainty range of -4.1 to 0.4 W/m^2 , mostly due to uncertainties in the indirect sulfate aerosol forcing. These estimates, which included sulfate, mineral dust, black and organic carbon from burning fossil fuel, and organic carbon from biomass burning, were based on *direct* calculations of the forcing using comprehensive radiative transfer and chemical transport models.

Recently, the IPCC estimates have been revised in two ways. First, Boucher and Haywood (2001), using different assumptions about the distribution of aerosol forcing within the IPCC range, ran a Monte-Carlo simulation to construct probability density functions (pdfs) for the radiative forcing, and particularly for the total aerosol forcing. They obtained a mean value of -1.6 W/m^2 and a 5% to 95% confidence interval of -3 W/m^2 to -0.5 W/m^2 , respectively. Their analysis showed that their pdfs are very sensitive to how they treated the distribution of the indirect sulfate aerosol forcing values. Second, Hansen and Sato (2001) revised their own estimates of the total aerosol forcing published in 1998 to obtain a total aerosol forcing with a mean of -0.6 W/m^2 and an uncertainty range of $-0.6 \pm 1.1 \text{ W/m}^2$. In the latter work the uncertainty range was skewed to the right toward less negative forcing, thereby allowing for the uncertainties due to carbonaceous aerosol and, particularly, due to the black carbon. Our synthesis of the data described above is presented on Fig. 1A.

The “*inverse*” estimations of the aerosol forcing have been based on comparison of multiple simulations by a simple climate model of the temperature change due to various sets of radiative forcings in comparison with the historical observational record. Most of these estimations included constraints on the direct and/or indirect sulfate forcing. Our pdf synthesis of all “*inverse*” estimations of the aerosol forcings (Wigley and Raper, 2001; Andronova and Schlesinger, 2001; Gregory et al, 2002; Knutti et al, 2002, Forest et al, 2002), together with the direct IPCC estimation of the sulfate forcing, are presented in Fig. 1B. From Fig. 1B it can be seen that the range of uncertainties for the inverse calculations are much smaller than for the direct calculations. Our own “*inverse*” calculations of the aerosol forcing (Andronova and Schlesinger (AS), 2001), based on the hemispheric version of our simple climate model, showed that adding any positive forcing, such as tropospheric ozone forcing, to the suite of radiative forcings moves the entire distribution of the sulfate radiative forcing to the left to reproduce the observed temperature change, thereby making it more severe (see Fig. 1C).

Aerosol forcing and climate sensitivity

Our calculations of the constrained aerosol forcing, presented in AS, were made simultaneously with the estimation of the values of the climate sensitivity. We constructed the pdf for climate sensitivity using 80,000 realizations of the climate noise, defined as a difference (residual) between the simulated (forced) temperature change and the observed temperature departure. A simple atmosphere/ocean model was used to drive temperature response to 16 different radiative forcing models. The radiative forcing models, used to simulate temperature anomalies, represented multiple combinations of individual forcings from greenhouse gases (including CO_2 , CFC-11, CFC-12, N_2O , CH_4), tropospheric ozone forcing, anthropogenic and natural sulfate aerosol forcing (direct and indirect), forcing due to variation in the solar constant and volcanic forcing. Realizations of the climate noise were constructed by using the bootstrap re-sampling method, which was applied 5000 times to the residual obtained for each radiative forcing model. Each individual value of the climate sensitivity was estimated by minimizing the root

mean square error between the simulated and observed temperatures. The results showed that in general the aerosol forcing estimations are inversely related to the estimations of the climate sensitivity -- with less negative aerosol forcing the climate sensitivity tends to be smaller. Thus, making aerosol forcing less negative (e.g., by inclusion of the carbonaceous aerosols) will reduce the climate sensitivity and will slightly reduce the uncertainties of the climate sensitivity estimations (see Fig. 1D).

Conclusion

To reduce the uncertainty in the estimation of climate sensitivity requires reducing the uncertainty in the radiative forcing, particularly by anthropogenic tropospheric aerosols. In spite of the progress made in understanding the mechanisms of the aerosol chemical interactions and radiative transfer, the uncertainties are large. Progress is needed in the direct estimation of the indirect aerosol forcing and the magnitude of the forcing by carbonaceous aerosol.

References

Andronova, N. G. and M. E. Schlesinger, 2001, Objective estimation of the probability density function for climate sensitivity. *J. Geophys. Res.*, 2001. 106(D19), p. 22,605-22,612.

Boucher, O. and J. Haywood, 2001, On summing the components of radiative forcing of climate change. *Climate Dynamics*, 18, p.297-302.

Forest, C. E., P. H. Stone, A. P. Sokolov, M. R. Allen and M. D. Webster, 2002, Quantifying Uncertainties in Climate System Properties with the Use of Recent Climate Observations. *Science*, 295, p. 113-117

Gregory, J. M., R. J. Stouffer, S. C. B. Raper, P. A. Stott and N. A. Rayner, 2002, An observationally based estimate of the climate sensitivity. *J. Climate*. 15(22), p. 3117-3121.

Hansen, J. E., and M. Sato, 2001, Trends of measured climate forcing agents, *Proc. Natl. Acad. Sci. U.S.A.*, 98, 14,778– 14,783.

IPCC (2001) *Climate Change 2001, the scientific basis, contributions of working group I to the third assessment report of the intergovernmental panel on climate change*. Houghton J. T., et al. (eds) Cambridge University Press, New York, USA.

Knutti, R., T. F. Stocker, F. Joos and G. K. Plattner, 2002, Constraints on radiative forcing and future climate change from observations and climate model ensembles. *Nature*, 416, p. 719-723.

Wigley, T. M. L. and S. C. B. Raper, 2001, Interpretation of High Projections for Global-Mean Warming, *Science*, 293, p 451-454.

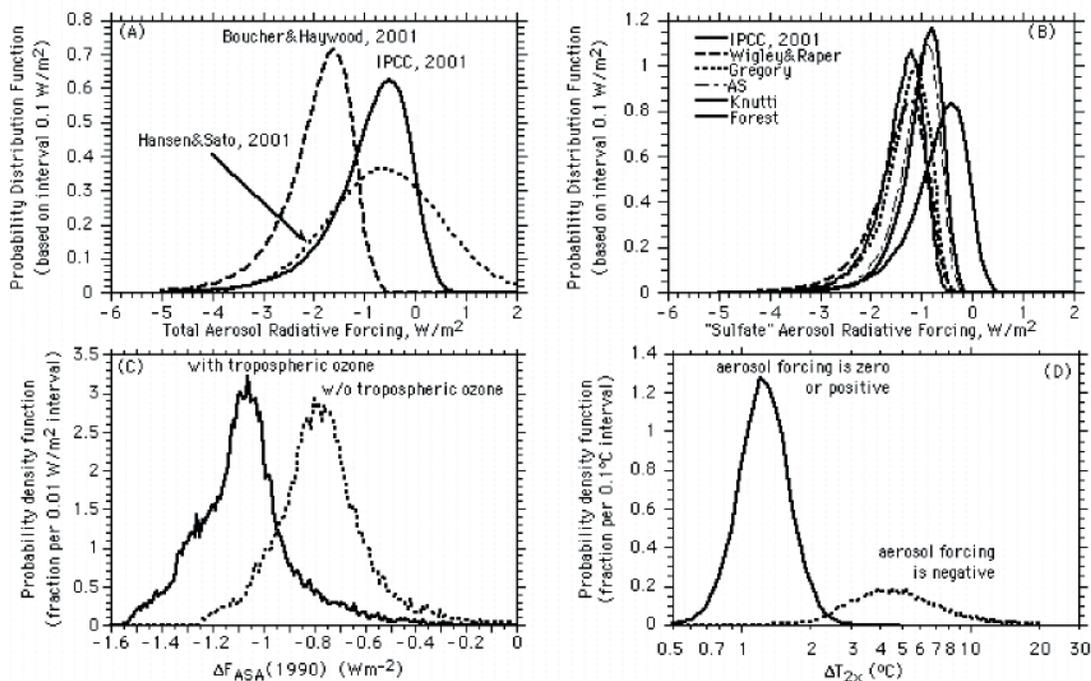


Figure 1. Probability distribution functions for (A) total aerosol forcing; (B) sulfate aerosol forcing; (C) the sulfate aerosol forcing for 1990 by Andronova and Schlesinger (2001) both with and without radiative forcing by tropospheric ozone; and (D) climate sensitivity for the two cases in (C) on a log scale.

Geographical Aspects of Feedback/Sensitivity

G.J. Boer

Canadian Centre for Climate Modelling and Analysis, Meteorological Service of Canada

Changes in the vertically integrated energy budget of the climate system are expressed as

$$dh'/dt = A' + R' = A' + g + f = A' + \Lambda T' + f \rightarrow 0 \text{ at equilibrium} \quad (1)$$

where $X' = X - X_0$ is the difference between a perturbed, X , and control, X_0 , climate quantity, dh'/dt energy storage, A' convergence of energy by transport processes and R' radiative flux into the column. R' is expressed as the sum of the radiative forcing $f(\lambda, \varphi, t)$ and the radiative feedback $g = \Lambda T'$, the product of the surface temperature change $T'(\lambda, \varphi, t)$ and the feedback parameter $\Lambda(\lambda, \varphi, t)$. The connection between global average forcing and temperature (indicated by angular brackets) follows as

$$\langle T' \rangle = \langle f - dh'/dt \rangle / \hat{\Lambda} = \hat{s} \langle f - dh'/dt \rangle \hat{s} \rightarrow \hat{s} \langle f \rangle \text{ at equilibrium} \quad (2)$$

where the global feedback parameter

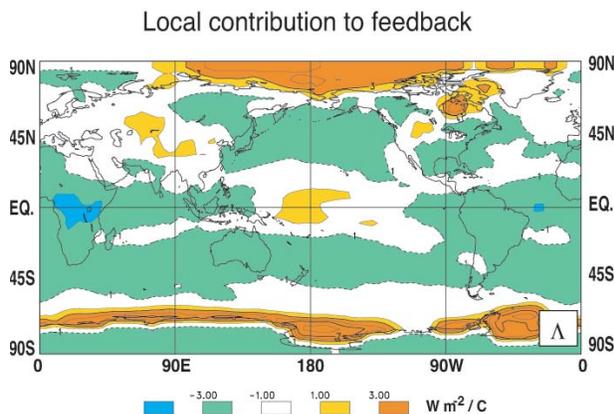
$\hat{\Lambda} = \langle \Lambda T' \rangle / \langle T' \rangle = -1 / \hat{s}$ is the temperature weighted average of the local feedback parameter $\Lambda(\lambda, \varphi, t)$ and is inversely proportional to the sensitivity parameter \hat{s} . For mixed layer ocean versions of coupled climate models, the sensitivity is found to differ by as much as a factor of two (e.g. Figure 9.18 of IPCC2001).

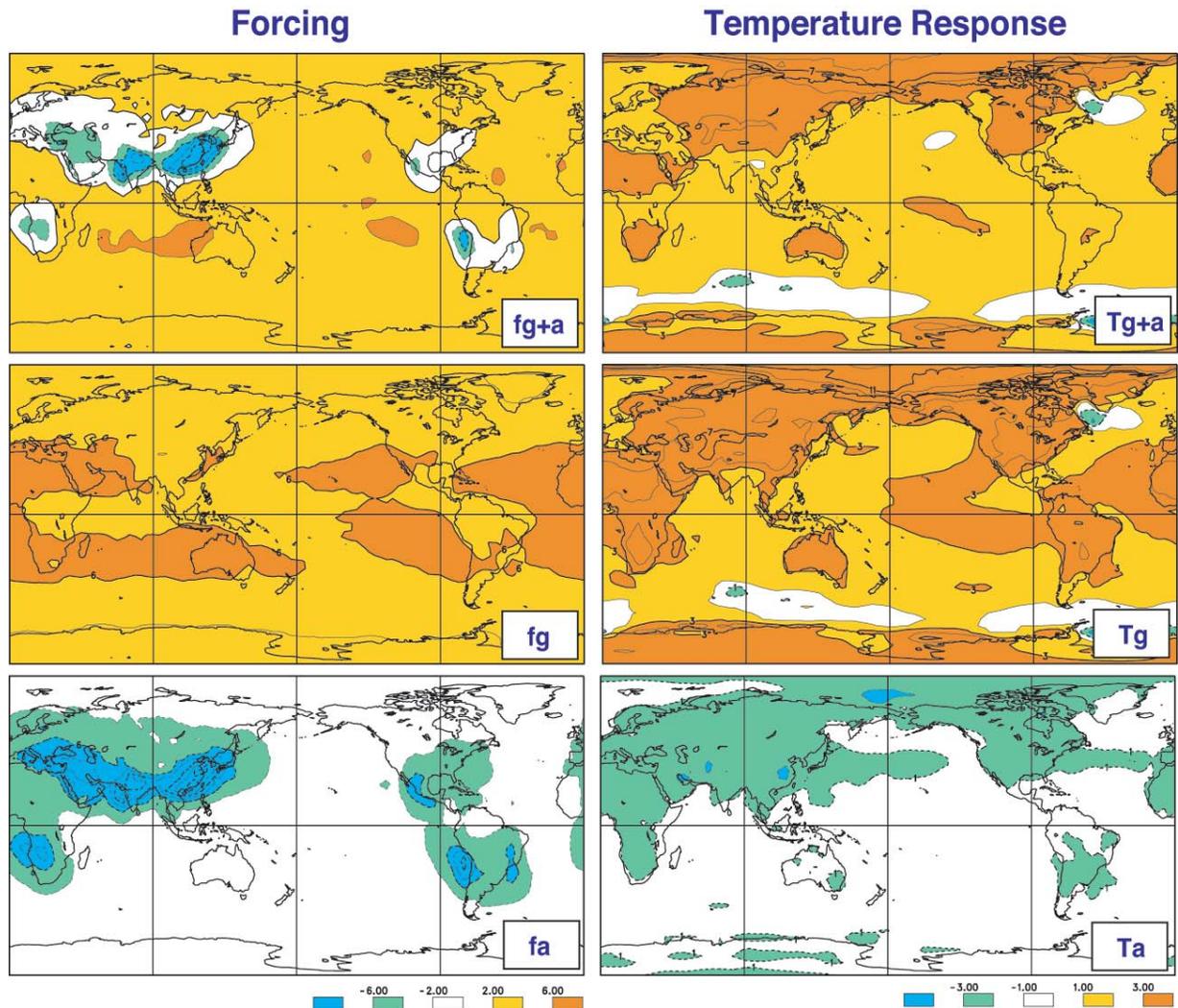
Boer and Yu (2003) argue that the geographical expression (1) gives a more basic representation of feedback/sensitivity than does the global representation (2). Figure 1 (below) displays the normalized local feedback parameter $\Lambda_l = \Lambda T' / \langle T' \rangle$, for which $\hat{\Lambda} = \langle \Lambda_l \rangle$, obtained from a near equilibrium simulation with the CCCma CGCM2 with greenhouse gas and aerosol concentrations stabilized at year 2050 values from the IS92a scenario. The high latitude regions of positive

feedback are associated with the surface ice/snow albedo feedback while those in the tropical Pacific depend on changes in the structure of moisture, temperature and cloudiness. Negative regions are associated with increased outgoing IR from the warmer surface and atmosphere and/or with or decreased solar associated with changes in cloud distribution and optical properties.

If $\Lambda(\lambda, \varphi, t)$ is a feature of the physical climate system which is independent of f and T' , at least to first order, then (1) indicates that the temperature response pattern $T'(\lambda, \varphi, t)$ need not resemble the forcing pattern $f(\lambda, \varphi, t)$ and may be spatially remote from it as seen in Figure 2 on the next page (where subscripts g and a indicate GHG and aerosol forcings and temperature responses). The forcing is redistributed by the transport and local feedbacks act to “localize” the temperature response. Forcing is amplified in the regions of positive feedback (e.g. at high latitudes) and damped in regions of negative feedback. Thus temperature response patterns are determined largely by the pattern of the feedbacks and only secondarily by the pattern of the forcing. Moreover, if A' is approximately linear in T' then (1) implies that forcing and response patterns add linearly (e.g. Forster et al., 2000). Finally, (2) indicates that global feedback $\hat{\Lambda}$ and sensitivity \hat{s} are not independent of the temperature and forcing patterns. Different forcing patterns with the same global mean $\langle f \rangle$ may give different global mean temperature responses. This is because forcing preferentially in regions of positive feedback gives a larger global mean temperature response than does forcing uniformly or forcing preferentially in regions of negative feedback. This “efficacy” of the forcing (e.g. Hansen and Nazarenko, 2004) arises from the local nature of the feedbacks as measured by $\Lambda(\lambda, \varphi, t)$.

The feedback g and the feedback parameter Λ can be decomposed into components as $g = g_S + g_L = g_A + g_C = g_{SA} + g_{SC} + g_{LA} + g_{LC}$ associated respectively with Solar, Longwave, clear-sky Atmosphere/surface and Cloud feedbacks. These components give some information on the different processes involved in the model response and may also form the basis for understanding and intercomparing the behaviour of different models as expressed by their radiative responses. A comparison of the global mean values of feedback components between two versions of the NCAR coupled model





(CCSM2.0.1, and CSM1.4) and the CCCma CGCM3 model, calculated for the last 10 years of a 50 year simulation forced with a switch-on 5% solar increase ($\langle f \rangle \approx 8 \text{ W m}^{-2}$), is shown in Table 1. The stronger negative feedback (hence lower sensitivity) of the NCAR models compared to CGCM3 is immediately apparent and the difference in solar cloud feedback Λ_{SC} is the main reason for this. Longwave cloud feedback is weakly negative and both long and shortwave clear-sky feedbacks are similar across models. The dependence of feedback/sensitivity on cloud treatment in models is reiterated.

In summary, the geographical distribution of feedback processes is a robust aspect of the climate system that explains a number of features of model behaviour and offers the possibility of intercomparing and better understanding model and system feedback/sensitivity.

References

- Boer, G.J. and B. Yu, 2003: Climate sensitivity and response. *Clim. Dyn.*, **20**, 415-429.
- Forster, P., M. Blackburn, R. Glover and K. Shine, 2000: An examination of climate sensitivity for idealized climate change experiments in an intermediate general circulation model. *Clim. Dyn.*, **16**, 833-849.
- Hansen, J. and L. Nazarenko, 2004: Soot climate forcing via snow and ice albedos. *PNAS*, **101**, 423-428.
- IPCC2001: *Climate change 2001, the scientific basis*. Cambridge University Press, 881pp

Table 1. Global averages of feedback parameter components for two versions of the NCAR CGCM and the CCCma CGCM3. Units are $\text{W m}^{-2} \text{ C}^{-1}$ (prepared by M. Stowasser of IPCC).

	$\langle \Lambda \rangle$	$\langle \Lambda_{SA} \rangle$	$\langle \Lambda_{SC} \rangle$	$\langle \Lambda_{LA} \rangle$	$\langle \Lambda_{LC} \rangle$
CSM1.4	-2.58	0.61	-0.97	-2.12	-0.10
CSM2.0	-2.30	0.74	-0.84	-2.12	-0.08
CGCM3	-1.29	0.68	0.33	-2.26	-0.04

What Can We Learn on Climate Sensitivity from Mid-Holocene Type Simulations?

Pascale Braconnot

IPSL/Laboratoire des Sciences du climat et de l'environnement, Orme des Merisiers, bât 701, 91191 Gif-sur-Yvette Cedex, France, email: pasb.braconnot@cea.fr.

Introduction

Past climate simulations offer an unique opportunity to test our ability to represent a climate different from the modern one. Here we consider GCM simulations of the mid-Holocene climate to assess the sensitivity of the climate system to a slightly different seasonality of the insolation forcing (incoming solar radiation at TOA). The mid-Holocene (6000 years ago; 6ka) period is one of the key periods of interest of the Paleoclimate Modeling Intercomparison Project (PMIP; (Joussaume et Taylor, 1995; PMIP, 2000) that was initiated to facilitate systematic comparisons between model simulations and between model and proxy data. PMIP also offers the opportunity to test the sensitivity of climate models, even though this aspect has not been fully developed during the first phase of the project (PMIP1) that considered only general circulation model of the atmosphere (AGCM). Coupled ocean-atmosphere models are now becoming a standard to study climate change, and several OAGCM simulations are now available for 6ka and will be considered in the following. Time periods, such as last glacial inception (about 115000 years ago, 115ka) or the Early Holocene also receive lots of attention from a wide modeling community. They will be considered in the second phase of the project (<http://www-lsce.cea.fr/pmip2>) during which feedbacks from ocean and vegetation as well as changes in climate variability will be analysed in more

details, in order to evaluate those coupled models that are use for future climate projection (Harrison et al., 2002). Some aspects of the 115 ka climate will also be considered at the end of this short report. Evaluation of model experiments depends on the existence of spatially explicit data sets that can be compared with output from the model simulations. PMIP plays a key role in triggering data synthesis for model-data comparisons (see Harrison 2000 for a synthesis).

Model response to mid-Holocene insolation

The major difference between mid-Holocene and present day climates comes from the Earth orbital configuration (Berger, 1978), and mainly from the position of the perihelion and from a larger obliquity. The seasonal cycle of the incoming solar radiation at the top of the atmosphere (insolation) was increased by about 5 % in the northern hemisphere and damped in the southern hemisphere. This seasonal change in insolation is the driving mechanism for the increase and the northward shift of summer precipitation over the African and Asian continents. During summer, temperature are warmer and precipitation larger over land whereas over the ocean precipitation decreases due to the larger moisture advection from the ocean to the continent (Figure 1a). In coupled experiment, SST can vary, and, in this case, some of the models produce local enhancement of precipitation over the warmer ocean waters (Figure 1b).

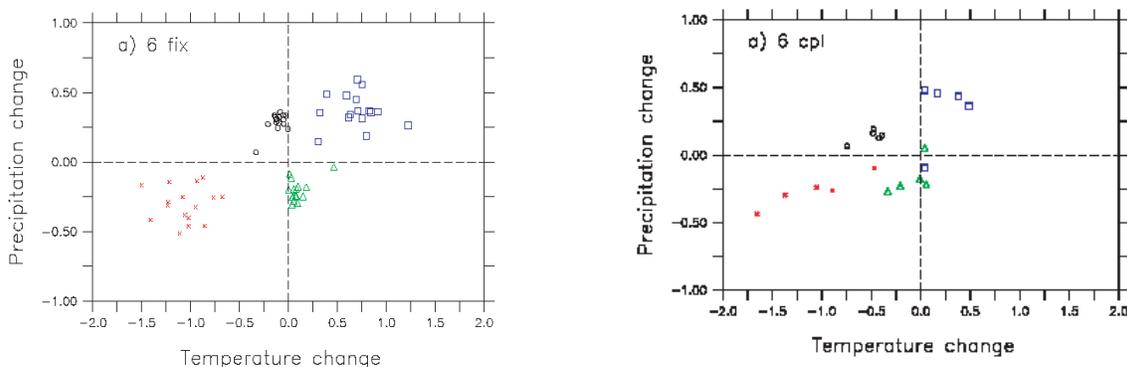


Figure 1. Change in 6ka tropical precipitation and temperature for a) PMIP simulations and b) coupled simulations averaged over the continent in winter (red dots), the ocean in winter (black dots), the continent in summer (blue squares) and the ocean in summer (green triangles). Each point represents a model results.

In winter, insolation is reduced and so is the temperature over land. The winter monsoon is enhanced and precipitation increases over the ocean. These changes are thus seasonal and correspond to shift in time and space of the major climate structures. There is nearly no change in annual mean. Change in the hydrology of the tropics is a major feedback, but cannot be analysed from a global point of view. The realisms of the simulated changes can be tested against different proxy indicators and synthesis map reconstructions. This type of analysis gives rise to the key diagnostics proposed in the TAR, where, a zonal mean estimate of the amount of moisture needed to produce the observed shift in the transition between desert and steppe in Sahara has been proposed. It allows evaluating how well models reproduce this feature. During the presentation an example using new pollen data reconstruction in middle and high latitudes will be presented.

Model differences in forcing and radiative feedbacks

The standard definition of climate sensitivity is not well adapted for this seasonal change. However it is possible from analyses of model-model differences to extract some information about model differences. The radiative forcing can be estimated from Hewitt and Michel (1996)'s definition of shortwave forcing for mid-Holocene: $SW_{\text{forcing}} = (1 - \alpha_{\text{ctrl}}) \Delta SW_i$. The rationale is that different climate models have different forcing because, even though we apply the same change in insolation (ΔSW_i) to all of them, the effective change in absorbed short wave radiation is a function of the planetary albedo of the control simulation (α_{ctrl}). The change in net SW at TOA thus results from the sum of this forcing and of internal feedbacks that alter cloud characteristics and surface albedo, and that are also responsible for differences between models results (Bonfils *et al.*, 2001; Joussaume *et al.*, 1999). The intensity of the change in precipitation in PMIP simulations was related to two factors of the control simulation. The first one is a parameter that account for the large-scale advection of moisture in the tropics and subgrid scale parameterisation of clouds and rain. The second one is the mean temperature of the control simulation (Braconnot *et al.* 2002). The change in the energetic of the atmospheric column (for dry atmosphere) was analysed for PMIP simulations following the ITCZ over west Africa and the region of maximum warming further north into the Sahara (Braconnot *et al.* 2000). Comparison between results obtained for 6 ka and last glacial maximum revealed that, although the contribution of the different fluxes to the atmospheric budget strongly vary from a model to an other, a same model produce similar characteristics

for 6ka and last glacial maximum which respond to very different forcing mechanisms (Figure 2).

Role of ocean and vegetation feedbacks

Ocean and vegetation feedbacks have been show to play a major role in the amplification of the climate response to the insolation forcing (i.e. Braconnot *et al* in press or Cane *et al*, 2004 for a synthesis). Only few simulations have considered simultaneously the ocean and vegetation feedback (Braconnot *et al.*, 1999). These results showed how, for the African monsoon, ocean and vegetation feedback respectively reinforce the inland advection of humidity and local recycling. These feedbacks will be illustrated for high latitudes from the analyses of Wohlfahrt *et al.* (2004). Changes in the land surface cover introduce an annual mean forcing on the climate resulting from changes in the surface albedo, surface roughness and roots profiles. This signal is superimposed on the seasonally varying mid-Holocene insolation forcing. In middle and high latitudes, the vegetation and ocean feedbacks respectively enhance the warming in Spring and Autumn. The synergy between the different feedbacks is for most seasons as large as the feedbacks themselves and contributes to translate the insolation seasonal forcing into a mean annual warming. As for the African monsoon the resulting climate change is larger than the sum of the individual contribution from ocean and vegetation feedbacks, stressing that their non-linear interactions play an important role and need to be considered in climate change experiments. However, even though the simulation including all the feedbacks is in better agreement with data when considering the forest extension over Europe, the match with biome data is reduced compared to AGCM or OAGCM over the mid-continental Eurasia where the drying become excessive. This point needs to be analysed from different simulations to determine if we miss a process, or a boundary condition such as lakes in the land surface scheme.

Feedback from atmospheric moisture transport

Coupled simulations of the mid-Holocene do not produce notable change in the annual mean overturning or thermohaline circulation. Changes in the ocean circulation have rather seasonal features (Braconnot *et al.*, 2000). The equator to pole heat transport is reduced during winter because the tropical regions are colder. In summer it helps to transfer the excess summer warming from the northern hemisphere to the southern hemisphere. The late warming of the surface ocean in response to the insolation forcing has been shown to enhance the inter-hemispheric contrast in temperature, which affects both the Indian and African monsoon

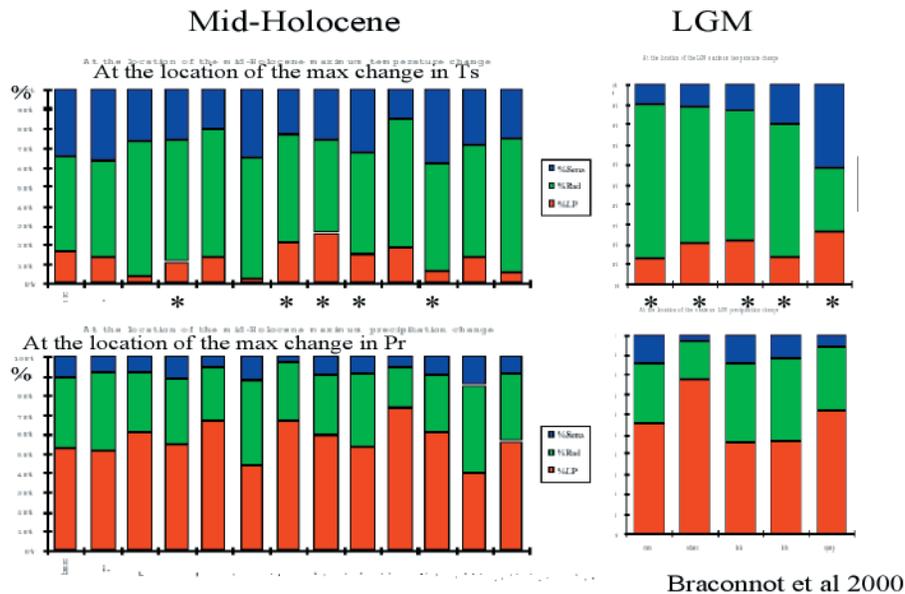


Figure 2. Partitioning of the latent heat release (LP), the radiative fluxes (Rad) and the sensible flux (Sens) in the energetic change of the atmospheric column ($Q = LP + Rad + Sens$) for mid-Holocene and last glacial maximum. Each bar represents a model result (PMIP simulations). Models that performed both 6ka and LGM simulations are marked with a star (same order from left to right). The bottom panel represents the annual mean within the ITCZ over west Africa. Values were computed for each month and then averaged. The top panel was obtained from similar calculation for the region of maximum warming located to the north of the ITCZ in the Sahara.

(zhao et al, in preparation). Interesting features are also found between simulations of 6ka and 115 ka for high latitudes when comparing the change in moisture transport and the feedback in high latitudes. Last glacial inception is a time during which high latitudes received less solar energy during summer and low latitude more energy during winter. The latitudinal pattern of insolation change as a function of latitude and month is roughly the reverse of the one found for 6ka. Using the IPSL coupled model, Khodri et al. (2001) showed that the increase in moisture transport from the equator to high latitudes was an important factor to trigger the last glacial inception. Some of the changes in the hydrological cycle of the Arctic show reverse behaviour between 115ka and 6ka. At 6ka (115ka), the sea ice cover is reduced (increased) during summer, the atmosphere transports less (more) moisture towards high latitudes, which reduces (increases) the river runoff discharge in the Arctic and prevents (favours) the building up of sea ice. In the case of 115ka these changes lead to colder summer and snow precipitation over the continents where ice sheet initiated (Khodri et al. 2001). Reconstruction of SST from proxy indicators suggests that the associated simulated change in the north-south SST gradient in the North Atlantic is consistent with reality.

Impact of the seasonal time scale between forcing and climate response

Other important feedbacks that may alter climate sensitivity arise from the timing between the climate forcing and the timing of the climate response. Recent results show that the late response of the tropical ocean to the mid-Holocene insolation results from the fact that the maximum change in insolation occurs after the summer solstice in July (Braconnot et al. 2003). Analysis of the feedback from snow and sea-ice in high northern latitude will be discussed from simulations where the perihelion has been respectively shifted to winter or to spring from 115 ka orbital configuration. Even with a large summer insolation, if the spring melting of snow and sea-ice does not occur early enough in the season the snow-albedo feedback is not effective and the high latitude climate is cold. These seasonal changes interfere with the ocean heat transport in the north Atlantic (Khodri et al. submitted), and may be one of the reason why marine data suggest that last glacial inception was already effective 120000 years ago (Corijo et al. 1999).

Conclusion

Even though the characteristics of climate change we present here do not represent analogue with what could

happen under increased green house gas concentration, they allow us to isolate major feedbacks and to understand the combination of scale between the change in mean climate, climate variability and extremes. Analysis of radiative changes in climates where the seasonal cycle is slightly different from the present one is also a way to better understand the present day seasonal cycle, and to test if the sensitivity of climate model is in agreement with proxy reconstruction of past environment. This work is entirely relevant to analyses for instance the climatic impact of future changes in regions of human emission of aerosols These forcing may be more effective in some seasons and contribute to changes in regional climates and climate teleconnections. Change in climate variability was not discussed here. One of the reasons is that, as for present day climate, climate models produce a wide range of possibilities and careful analyses are now required to define which part of the signal is robust and which part is model dependent. This will be one of the major challenges for PMIP2 and both the modelling and data communities in the coming years.

References

- Berger, A., 1978: Long-term variations of caloric solar radiation resulting from the earth's orbital elements. *Quaternary Research*, 9, 139-167.
- Bonfils, L., N. de Noblet-Ducoure, P. Braconnot, et S. Joussaume, 2001: Hot desert albedo and climate change: Mid-Holocene monsoon in North Africa. *Journal of Climate*, 14, 3724-3737.
- Braconnot, P. et O. Marti, 2003: Impact of precession on monsoon characteristics from coupled ocean atmosphere experiments: changes in Indian monsoon and Indian ocean climatology. *Marine Geology*, 201, 23-34.
- Braconnot, P., M. F. Loutre, B. Dong, S. Joussaume, et P. Valdes, 2002: How the simulated change in monsoon at 6 ka BP is related to the simulation of the modern climate: results from the Paleoclimate Modeling Intercomparison Project. *Climate Dynamics*, 19, 107-121.
- Braconnot, P., O. Marti, S. Joussaume, et Y. Leclainche, 2000: Ocean feedback in response to 6 kyr BP insolation. *Journal of Climate*, 13, 1537-1553.
- Braconnot, P., S. Harrison, J. Joussaume, C. Hewitt, A. Kitoh, J. Kutzbach, Z. Liu, B. L. Otto-Bleisner, J. Syktus, et S. L. Weber, press: Evaluation of coupled ocean-atmosphere simulations of the mid-Holocene. Kluwer.
- Braconnot, P., S. Joussaume, N. de Noblet, et G. Ramstein, 2000: Mid-holocene and Last Glacial Maximum African monsoon changes as simulated within the Paleoclimate Modelling Intercomparison Project. *Global and Planetary Change*, 26, 51-66.
- Braconnot, P., S. Joussaume, O. Marti, et N. de Noblet, 1999: Synergistic feedbacks from ocean and vegetation on the African monsoon response to mid-Holocene insolation. *Geophys. Res. Lett.*, 26, 2481-2484.
- Cane, M., P. Braconnot, A. Clement, H. Gildor, S. Joussaume, M. Khodri, D. Paillard, S. Tett, E. Zorita, 2004 : Progress in Paleoclimate Modeling, Special issue of the CLIVAR 2004 conference, smi.
- Cortijo, E. et al., 1999: Changes in meridional temperature and salinity gradients in the North Atlantic Ocean (308-728N) during the last interglacial period. *Paleoceanography* 14, 23-33).
- Harrison, S., P. Braconnot, C. Hewitt, et R. J. Stouffer, 2002: Fourth international workshop of The Palaeoclimate Modelling Intercomparison Project (PMIP): launching PMIP Phase II. EOS.
- Harrison, S., 2000: Palaeoenvironmental data set and model evaluation in PMIP. Vol. WCRP-111, WMO/TD-No. 1007, Paleoclimate Modeling Intercomparison Project (PMIP), proceedings of the Third PMIP workshop.
- Hewitt, C. D. et J. F. B. Mitchell, 1996: GCM simulations of the climate of 6 kyr BP: Mean changes and interdecadal variability. *Journal of Climate*, 9, 3505-3529.
- Joussaume, S. et K. E. Taylor, 1995: Status of the Paleoclimate Modeling Intercomparison Project. in Proceedings of the first international AMIP scientific conference, WCRP-92, Monterey, USA, 425-430.
- Joussaume, S., et al., 1999: Monsoon changes for 6000 years ago: Results of 18 simulations from the Paleoclimate Modeling Intercomparison Project (PMIP). *Geophysical Research Letters*, 26, 859-862.
- Khodri, M., Y. Leclainche, G. Ramstein, P. Braconnot, O. Marti, et E. Cortijo, 2001: Simulating the amplification of orbital forcing by ocean feedbacks in the last glaciation. *Nature*, 410, 570-574.
- PMIP, 2000: Paleoclimate Modeling Intercomparison Project (PMIP), proceedings of the third PMIP Workshop. WCRP-111, WMO/TD-No. 1007, 271 pp.
- Wohlfahrt, J., S. P. Harrison, et P. Braconnot, 2004: Synergistic feedbacks between ocean and vegetation on mid- and high-latitude climates during the mid-Holocene. *Climate Dynamics*, 22, 223-238.

Annex 2: Extended Abstracts

Diagnosing Climate Sensitivity and Feedbacks in GFDL's AM2 Model

A. J. Broccoli¹, B. J. Soden², R. Hemler², R. Gudgel², T. Knutson² and S. Weber¹

¹ *Dept. of Environmental Sciences, Rutgers University, New Brunswick, NJ, USA*

² *NOAA/Geophysical Fluid Dynamics Laboratory, Princeton, NJ, USA*

The development of GFDL's new atmospheric general circulation model, known as AM2, provides an opportunity for a detailed evaluation of the evolution of the model's climate sensitivity and the associated feedback mechanisms. Early in the development process, the climate sensitivity to a doubling of CO₂ was relatively high, ranging from 3.9-4.6 K in early versions of AM2. A marked decrease in sensitivity accompanied a change in the model's turbulence parameterization, such that the 2xCO₂ sensitivity of the latest version is ~2.8 K. Changes in the strength of the cloud feedback are primary responsible for the decrease in sensitivity.

An analysis of individual feedback strengths indicates that the surface albedo and net cloud feedbacks are generally in the range of models surveyed previously. The water vapor and lapse rate feedbacks are both relatively large compared to other models, but their effects largely offset because of their opposing signs.

A comparison of diagnosed feedbacks from atmosphere-slab ocean model simulations and prescribed sea surface temperature perturbation ("Cess") simulations reveals substantial differences in cloud feedback strengths. Differences in the surface warming patterns are believed to be primarily responsible.

The Cess experiments were also used to compare were changes in cloud radiative forcing with cloud feedbacks diagnosed offline using the partial radiative perturbation method. These comparisons indicated that changes in cloud radiative forcing are not equivalent to cloud feedback strength, with the former exhibiting a negative bias. If this bias is accounted for, virtually all climate models evaluated to date would have neutral or positive net cloud feedback, leaving thermal damping (i.e., the "Planck feedback," or combined surface temperature and lapse rate feedbacks) as the only substantial negative feedback.

Annex 2: Extended Abstracts

The 1997/98 El Niño: A Test for Climate Models

Robert D. Cess¹, Riyu Lu², Buwen Dong³, and Gerald L. Potter⁴

¹ Marine Sciences Research Center, State University of New York at Stony Brook, Stony Brook, New York, 11790, USA

² Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing 100080, China

³ Hadley Centre for Climate Prediction and Research, Met Office, FitzRoy Road, Exeter EX1 3PB, UK

⁴ Program for Climate Model Diagnosis and Intercomparison, Lawrence Livermore National Laboratory, Livermore, California, 94550, USA

Cloud-climate interactions comprise one of the greatest uncertainties in attempting to model climate change using general circulation models (GCMs), and there is a need to devise ways of testing such interactions within models. If a GCM is to properly portray long-term climate change, it in turn must replicate cloud changes associated with events occurring on shorter time scales. Recently *Cess et al.* [2001] demonstrated that the lack of a zonal SST gradient in the tropical Pacific Ocean during the 1997/98 El Niño caused a collapse of the Walker circulation together with enhanced upward motion over the eastern Pacific. This in turn resulted, on average, in lower clouds in the western portion of the tropical Pacific and higher clouds in the eastern portion. The data used to determine these west/east cloud changes consisted of cloud profiling measurements made by the Stratospheric Aerosol and Gas Experiment (SAGE) II, together with cloud-radiative forcing (CRF) as determined for a 5-yr period (1985-1989) and for the first 4 months of 1998 by the Clouds and the Earth's Radiant Energy System (CERES) instrument on the Tropical Rainfall Measuring Mission (TRMM) Satellite. As emphasized by *Cess et al.* [2001], these cloud structure changes should serve as a useful test of a GCM, and in the present study we demonstrate this by applying this test to Version 3 of the Hadley Centre Atmospheric Model (HadAM3) [*Pope et al.*, 2000].

As in *Cess et al.* [2001], we employ zonal winds from the National Environmental Prediction Center/National Center for Atmospheric Research (NCEP/NCAR) reanalysis [*Kistler et al.* [2001]]. The ERBE/CERES data are likewise the same as used by *Cess et al.* [2001] and for the same western and eastern study regions. The ERBE TOA measurements are for the first four months (JFMA) of 1985-1989 together with CERES measurements for the first four months of 1998 that represent the strongest period of the 1997/98 El Niño for which CERES data are available (CERES/TRMM data cover January through August 1998).

Two experiments have been performed with HadAM3. One consists of a control, or “normal year”, experiment in which climatological sea-surface

temperatures (SSTs), averaged over 30 years from 1961 to 1990 [*Smith and Reynolds*, 1998], were prescribed. The other experiment, the “El Niño” experiment, used observed SSTs from June 1, 1996 to August 31, 1998 [*Reynolds and Smith*, 1994]. These two integrations were different only in their initial conditions, which were taken from the end of spin-up integrations. For the normal-year integration, the spin-up with climatological SSTs lasted 1.5 years, after which the control experiment was run for one year beginning on December 1. For the El Niño experiment the spin-up, using observed SSTs, was from June 1, 1996 to December 31, 1996, after which the El Niño experiment was integrated from January 1, 1997 to August 31, 1998. Each experiment consists of the average of an ensemble of 10 runs.

A major change in the atmospheric circulation over the tropical Pacific, as caused by the 1997/98 El Niño, was the virtual collapse of the Walker circulation [*Bell et al.*, 1999; *Cess et al.*, 2001], and this in turn caused the west/east changes in cloud structure noted by *Cess et al.* [2001] employing both SAGE II cloud profiling and ERBE/CERES TOA radiometric measurements. This collapse of the Walker circulation is demonstrated by comparing the NCEP/NCAR zonal winds in Figure 1c (El Niño) to those in Fig. 1a (normal year). The upward branch of the Walker circulation is centered between 140°E and 160°E (Figure 1a), which corresponds to strong convection that produces high clouds in the western region. The downward branch is located to the east of 160°W, with subsidence air suppressing deep convection and producing shallow stratus and stratocumulus clouds. The Walker circulation has all but ceased in 1998 (Figure 1c), and the westerlies east of the dateline in the upper troposphere have disappeared. This collapse of the Walker circulation is directly associated with the lack of a zonal SST gradient, and further discussion of this is given in *Cess et al.* [2001]. Thus a crucial first step in using the 1997/98 El Niño as a test of a GCM is to test if the GCM produces a similar collapse of the Walker circulation, and Figure 1 demonstrates that HadAM3 does a remarkably good job in this regard. This agreement of the model with the NCEP/NCAR

horizontal winds is consistent with a related study [Allan et al., 2002] who found agreement between HadAM3 and NCEP/NCAR 500-hPa vertical velocities.

This study has demonstrated, through the use of HadAM3, the utility of using the strong 1997/98 El Niño as one means of testing cloud-climate interactions within a GCM. A crucial first step is to determine if a GCM produces the observed collapse of the Walker circulation, and HadAM3 did a remarkable job of passing this test. The second phase of the test is to compare the model to ERBE/CERES TOA measurements, and the model did produce trends similar to those inferred from the satellite measurements. On average, cloud altitudes decreased over the western Pacific and increased over the eastern Pacific, both caused by the collapse of the model’s Walker circulation. We feel this is a useful test to apply to other GCMs.

References

Allan, R. P., A. Slingo, and M. A. Ringler (2002), Influence of dynamics on the changes in tropical cloud radiative forcing during the 1998 El Niño, *J. Climate*, 15, 1979-1986.

Bell, G. D., M. S. Halpert, C. F. Ropelewski, V. E. Kousky, A. V. Douglas, R. C. Schnell, and M. E. Gelman (1999), Climate assessment for 1998, *Bull. Am. Meteorol. Soc.*, 80, S1-S48.

Cess, R. D., M. H. Zhang, P.-H. Wang, and B. A. Wielicki (2001), Cloud structure anomalies over the tropical Pacific during the 1997/98 El Niño (2001), *Geophys. Res. Lett.*, 28, 4547-4550.

Pope, V. D., M. L. Gallani, P. R. Roundtree, and R. A. Stratton (2000), The impact of new physical parameterizations in the Hadley Centre climate model–HadAM3, *Climate Dyn.*, 16, 123-126.

Kistler, R. et al (2001), The NCEP-NCAR 50-year reanalysis: Monthly means CD-ROM and documentation, *Bull. Amer. Meteor. Soc.*, 82, 247-267.

Reynolds, R. W., and T. M. Smith (1994), Improved global sea surface temperature analyses using optimum interpolation, *J. Climate*, 7, 929-948.

Smith T. M., and R. W. Reynolds (1998), A high-resolution global sea surface temperature climatology for the 1961-90 base period, *J. Climate*, 11, 3320-3323.

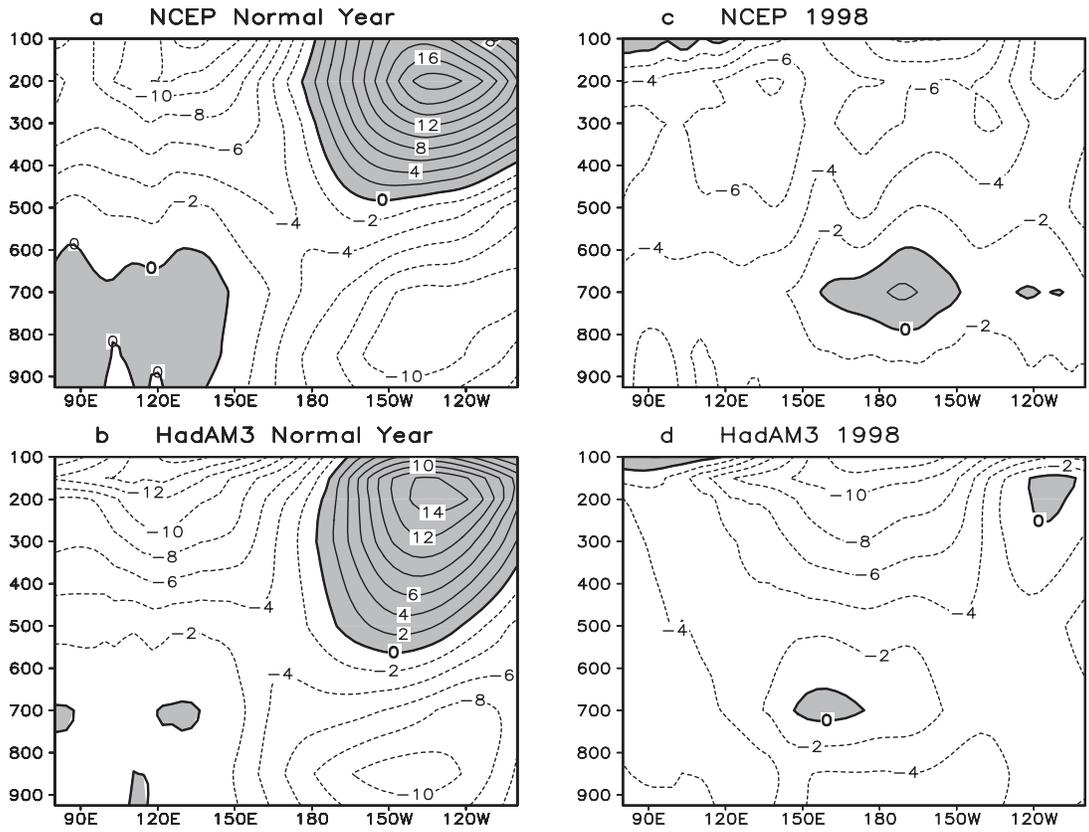


Figure 1. Pressure–longitude cross sections of zonal wind (m/s) averaged from 5°S to 5°N and for the first four months (JFMA) of each year. (a) NCEP normal year. (b) HadAM3 normal year. (c) NCEP 1998. (d) HadAM 1998.

Understanding and Comparing Climate Feedbacks in Models

Rob Colman

Bureau of Meteorology Research Centre, Melbourne, Australia

Introduction

Uncertainty in the strengths of important climate feedbacks underlies much of the uncertainty in the expected magnitude of future climate change (IPCC, 2001). Diagnosing, understanding and intercomparing climate model feedbacks are important steps towards reducing this uncertainty. This study first examines aspects of the time and space structure of the radiative perturbations contributing to the climate feedbacks in one particular model. These are the vertical and meridional structure of water vapour feedback and the annual cycle of feedback contributions in the BMRC model coupled to a mixed layer ocean, and forced with a doubling of CO₂. A basic comparison between published results of feedbacks, and those of the BMRC model is also presented.

The present method of calculation of the feedbacks is a variant on the method proposed by Wetherald and Manabe (1988). In this technique, top of atmosphere (TOA) radiative changes between the control and perturbed model climates associated with changes to a particular parameter (e.g. water vapour) are diagnosed by ‘off-line’ running of the model radiation code, after substituting the parameter of interest from the perturbed climate. Details of the approach are given in Colman (2003b).

Structure of water vapour feedback in a model

A common finding with climate models is that the relative humidity (RH) stays roughly constant under climate change. This being the case, the pattern of temperature change in the troposphere largely controls the degree of moistening in each location, and therefore its contribution to the water vapour feedback. Hence, amplified upper tropospheric warming associated with lapse rate changes may be expected to produce a local maximum in water vapour change. The contribution of this component to the overall strength of the water vapour feedback, however, has not previously been quantified. In this study, using equilibrium 1 → 2×CO₂ changes from the BMRC model, three perturbations are applied on top of the control model climate: (1) RH changes, (2) temperature changes uniform with height, of the magnitude of the surface temperature change (with fixed RH), and (3) tropospheric temperature changes corresponding to the change in lapse rate at

each location, but with surface temperatures held fixed (also with fixed RH).

It is found that only a small global contribution (~2%) to the total water vapour feedback results from changes to RH. The fixed lapse rate, fixed RH, temperature perturbation is responsible for most of the feedback strength, with maximum contributions from the upper troposphere. Because of the polar amplification of warming, high latitudes play a significantly greater role than for the uniform temperature perturbation analysed in Held and Soden (2000). The fractional contribution to total water vapour feedback evaluated from fixed RH lapse rate changes in the model is around 20-25%, and is responsible for greatly increasing the importance of the tropics compared with high latitudes in determining the final strength of the water vapour feedback. Furthermore, in the BMRC model, the feedbacks resulting from lapse rate changes almost exactly compensate between temperature and water vapour, with a very similar structures in the tropics.

Seasonal cycle of contributions to feedbacks

Are the radiative contributions to the global feedbacks fairly constant over the year or do they vary significantly with the seasons? To address this question, monthly means of the TOA radiative perturbations (which together comprise the annual climate feedbacks) were extracted to produce a mean annual cycle for the model. Results are shown in Fig. 1.

For the long wave (LW) feedback, the associated radiative changes are, indeed, fairly constant throughout the year, particularly those due to water vapour and cloud feedbacks. Contributions to the total short wave (SW) feedback, on the other hand, vary by a factor of three, from a maximum in July to a minimum in the November to February period. Of these, contributions to both albedo and cloud show large seasonal variations, with the water vapour term close to constant. Radiative perturbations contributing to albedo feedback, as would be expected, vary in strength with snow and sea ice retreat which occurs at different latitudes and in different months. SW radiative perturbations from cloud changes vary markedly, changing sign between opposite seasons. The main source of this seasonal variation occurs from cloud amount changes. In the BMRC model, at least, this represents yet another layer of ‘offsetting’ in the

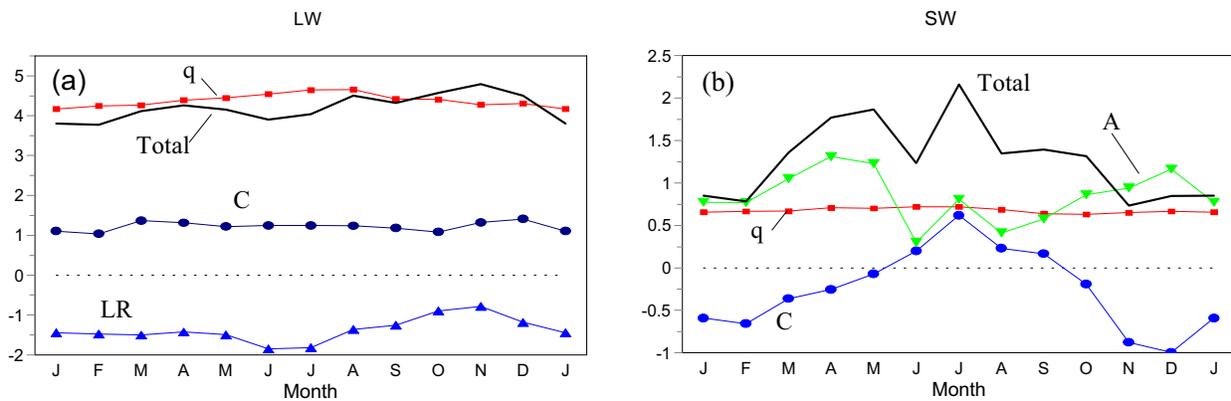


Figure 1. Globally averaged TOA radiative perturbations (for an equilibrium $1 \rightarrow 2\times\text{CO}_2$ climate change) as a function of month, due to feedbacks in the (a) LW and (b) SW. Units: Wm^{-2} . Shown are perturbations associated with the following feedbacks: 'q', water vapour; 'LR', lapse rate; 'C', cloud; 'A', albedo, as well as the total (see Colman, 2003b).

determination of cloud feedback: seasonal radiative perturbations of opposite sign contribute to the annual mean (which is close to zero). This also suggests caution must be exercised, for example, in the comparison of cloud feedbacks by 'traditional' perturbed SST, fixed season, experiments (e.g. Cess et al, 1990): results for particular models could vary dramatically depending upon the season chosen.

A comparison with model published feedback data

The final part of this paper considers a comparison of BMRC feedbacks with those from published results of 'offline' feedback calculations for GCMs with mixed layer oceans performing $2\times\text{CO}_2$ and solar perturbation experiments. Results (shown in Fig. 2) are from calculations using variants of the 'offline' TOA radiation perturbation technique from Wetherald and Manabe (1988) (and do not include the simplified model calculations quoted in Colman, 2003a). Also shown on this plot are three versions of the BMRC model (points connected by lines).

All feedbacks show substantial inter-model spread. Interestingly, the BMRC model versions span much of the model range for both water vapour and lapse rate feedbacks, and vary significantly in cloud feedback (but show less variation in the albedo feedback). The low water vapour feedback version of the model was an older 9 level model, with a relative humidity based cloud scheme and a 180 hPa imposed cloud ceiling. An offsetting of the water vapour and lapse rate feedbacks is apparent from the closer clustering of the BMRC results for the combined feedbacks (shown on the right). Indeed, a negative correlation appears to occur between water vapour feedback and lapse rate feedbacks across many models, and also within a single model (the BMRC2 model) under a broad range of climate forcing ($\frac{1}{2} \rightarrow 4\times\text{CO}_2$). There is also evidence for a

negative correlation between LW and SW components of the cloud feedback in the models that report them. The mean values of the feedbacks are comparable with results derived from the Cess et al. (1990) model intercomparison, when allowances are made for differences in diagnostic method and experimental paradigm.

Conclusions

The range of climate model sensitivity remains large despite marked progress in the development of climate models of the last two decades. Understanding the reasons for the spread in climate sensitivity is an important ongoing research goal, and a multipronged approach is vital. One important element is that, where possible, the values of feedbacks need to be diagnosed and compared between different models. Note that an updated climate feedback intercomparison, as well as further simplified intercomparisons are planned as part of the Cloud Feedback Model Intercomparison Project (see <http://www.cf mip.net/>). A second important element involves the detailed investigation of climate feedbacks in a single model, so as to understand the critical regions and physical processes which go to determine the strength of the individual feedbacks, as well as relationships between the feedbacks. This is important because intercomparisons between models cannot explore individual feedbacks to the same depth. Aspects of these two approaches are addressed in this study. Other vital steps include the direct comparison of model processes with observations, and, where possible, model large-scale feedbacks with observational surrogates (e.g. Soden et al., 2002).

References

Cess, R.D. and co-authors (1990) Intercomparison and interpretation of cloud-climate feedback processes

in nineteen atmospheric general circulation Models. *J Geophys Res* 95, 16601-16615.

Colman, R.A. (2003a) A comparison of climate feedbacks in general circulation models. *Climate Dynamics*, 20: 865-873.

Colman, R.A. (2003b) Seasonal Contributions to Climate Feedbacks. *Climate Dynamics*, 20, 825-841.

Held, I.M. and Soden, B.J. (2000) Water vapour feedback and global warming. *Ann. Rev. Energy Env.*: 25, 441- 475

IPCC (2001) *Climate Change 2001: The Scientific Basis*. Houghton JT, Ding, Y, Griggs, DJ, Noguer, M, van der Linden, PJ, Dai, X, Maskell, K and Johnson, CA (eds). Cambridge U Press, Cambridge, UK, 881pp.

Soden, B.J, Wetherald, R.T., Stenchikov, G.L. and Robok, A. (2002) Global cooling after the eruption of Mount Pinatubo: a test of climate feedback by water vapour. *Science*, 296, 727- 730.

Wetherald, R.T. and Manabe, S. (1988) Cloud feedback processes in a GCM. *J Atmos Sci* 45, 1397-1415.

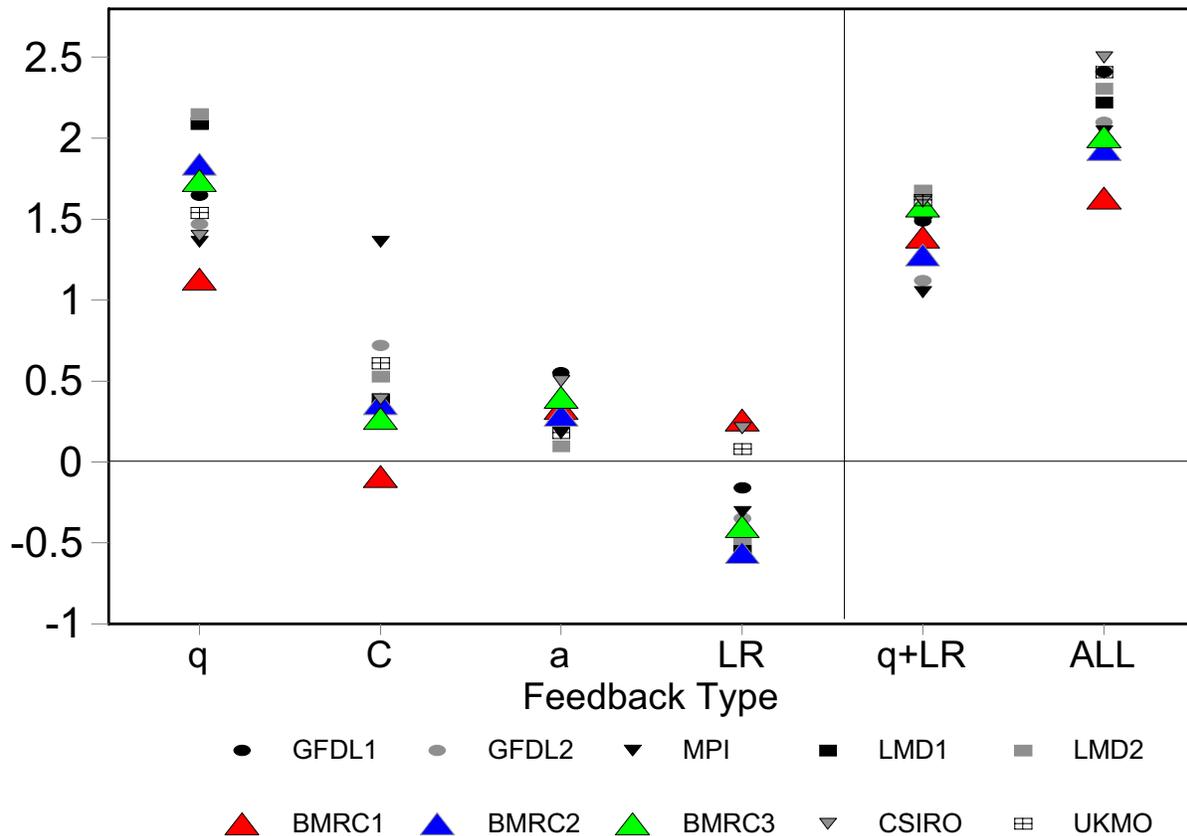


Figure 2. Global climate feedbacks (in $Wm^{-2}K^{-1}$) for a range of climate models (see Colman, 2003a). Positive values indicate a climate warming (i.e. increased downwards TOA radiation with increased surface temperature). Feedbacks are: Q (water vapour), C (clouds), A (albedo) and LR (lapse rate). Also shown are Q+LR (water vapour plus lapse rate) and ALL (sum of all feedbacks). All results are quoted from studies using 3D radiation perturbation methods. Coloured triangles denote results from three different versions of the BMRC model.

Annex 2: Extended Abstracts

Observational Constraints and Prior Assumptions on Climate Sensitivity

David Frame and Myles Allen

Atmospheric, Oceanic and Planetary Physics, University of Oxford, Clarendon Laboratory, Parks Road, Oxford OX1 3PU, UK

Abstract: The commitment to “stabilise atmospheric greenhouse gas concentrations” below a “dangerous” level appears to require an objective range of uncertainty in equilibrium climate sensitivity (ECS). We find that any such range depends critically on subjective prior assumptions regarding the distribution of ECS before any physical or observational constraints are applied. This is particularly true of the likelihood assigned to the high sensitivity values that are most relevant to any assessment of whether a given stabilisation level is considered “dangerous”. Indefinite stabilisation is, however, an unrealistic scenario. We argue that objective ranges of uncertainty on scientific inputs to most policy-relevant questions require a constraint on the transient climate response (TCR), not on ECS. These inputs include predicted transient 1990-2100 warming under the full range of emissions scenarios, peak warming assuming a realistic decline in greenhouse gas levels after the 21st century and the warming commitment due to greenhouse gas emissions to date. TCR is well constrained by the warming attributable to greenhouse gases over the 20th century, providing a robust and objective constraint on policy-relevant questions, bypassing the need for an objective range on ECS. The inherent difficulty in placing objective upper bounds on ECS means that policies aiming to avoid dangerous anthropogenic interference in the climate system may need to be formulated in terms of a peak concentration and subsequent minimum rate of decline rather than indefinite stabilisation.

All studies attempting to constrain sensitivity objectively can be couched in a Bayesian formulation in which the “posterior distribution”, $P(ECS|data)$, is proportional to the “likelihood” of the data assuming a given value of ECS (specifically, the likelihood of obtaining a model consistent with observations and with this particular value of ECS, $P(data|ECS)$) multiplied by the “prior”, $P(ECS)$. Recent studies [1,2,3] using the combination of observed warming over the 20th century and estimates of the recent rise in ocean heat content [4] to constrain ECS have arrived at highly asymmetric posterior distributions, assigning a relatively high likelihood to high sensitivities. The reason is the well-documented non-linear relationship between ECS and observed greenhouse-induced warming under almost any transient forcing scenario [5,6]: sensitive climates

take longer to reach equilibrium, making it difficult to place an upper bound on ECS based on transient warming to date.

This point is illustrated in the figure (a) on the next page, which shows the relationship between ECS (colour coding), warming attributable to greenhouse gases over the 20th century (vertical axis) and effective heat capacity of the atmosphere/ocean system (horizontal axis) in a simple climate model driven with observed greenhouse gas forcing. Although ECS has been sampled uniformly over the range 0.17 to 20K, the diamonds are not distributed uniformly in the vertical because of the non-linear relationship between ECS and transient response.

The black ellipse in figure (a) encloses the region consistent, at the 95% level, with the combination of warming attributable to greenhouse gases over the 20th century [7] and the effective heat capacity implied by observations of surface and subsurface temperature change over the 1957-1997 period [4]. The ellipse is based on published estimates of uncertainty in greenhouse and other forcings as well as surface and subsurface temperature trends: although contentious, these do not affect our basic point. If, following refs. [1-3], we assume a uniform (“neutral”) prior in ECS and heat capacity (all diamonds equally likely before the comparison with observations is made) and subsequently weight the diamonds by the likelihood of the observations given each particular combination of ECS and heat capacity, the resulting distribution $P(ECS|data)$ is shown as the red curve in figure (c). The “fat tail” extending out to high sensitivities arises from the fact that the ellipse in figure (a) includes a region (upper right quadrant) where the prior density, $P(ECS)$, is very high in the space defined by the observable quantities used to constrain the forecast.

A very different approach is taken by ref. [8], who perturb parameters in a climate model, weight the members of the resulting “perturbed-physics ensemble” by a measure of their similarity to observed present-day climatology (as opposed to recent climate change), and infer a distribution for ECS from the weighted ensemble. They sample parameters to provide a uniform distribution of the atmospheric feedback parameter, λ , which is inversely proportional to ECS, before the constraints of the perturbed-physics ensemble and

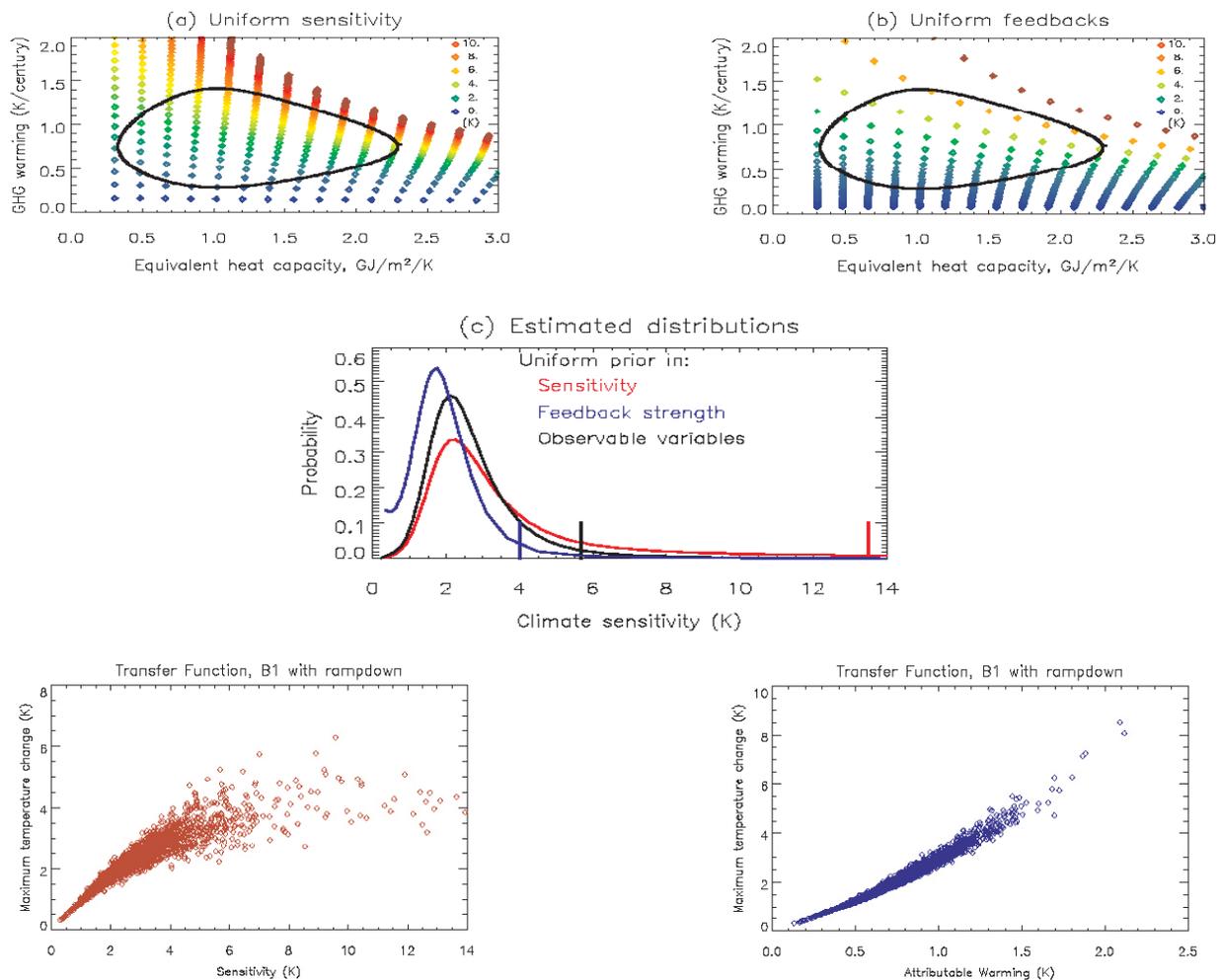


Figure. (a) and (b): Relationship between climate sensitivity (colours), effective ocean heat capacity and 20th century warming attributed to changes in greenhouse gases. Diamonds show energy balance model simulation results based on (a) uniform sampling of climate sensitivity, S , and (b) uniform sampling of feedback strength, or $1/S$. Black contours enclose the region consistent with observations at the 5% level. Panel (c) shows distributions of climate sensitivity based on these observations, assuming a uniform prior in sensitivity (red), in feedback strength (blue) and in attributable warming and heat capacity (black), with vertical lines showing 95th percentiles. Panel (d) shows the relationship between climate sensitivity and peak warming under a B1 scenario followed by a ramp-down over the 22nd and 23rd centuries, while (e) shows the corresponding relationship between past attributable warming and peak warming under a B1-plus-rampdown scenario.

climatology are applied. We can do the same with our simple model, and the result is shown in figure (b). Because we have now sampled $1/(ECS)$ uniformly in place of ECS, diamonds cluster towards the horizontal axis. After applying the likelihood-weighting, the resulting posterior distribution for ECS is shown by the blue curve in figure (c), which suggest a much (factor of 3) lower 95th percentile for ECS. Which upper bound is correct?

What we are seeing is a version of "Bertrand's paradox": suppose we have a mixture of water and wine and know only that the ratio of water to wine is somewhere between 2:1 and 1:1. For each unit of wine, we have between 1 and 2 units of water, so in

the absence of further information, our "best guess" is 1.5 units of water, implying a ratio of 3:2. But for each unit of water, we have between 0.5 and 1 unit of wine, so a "best guess" of 0.75, implying a ratio of 4:3. The difference arises from the fact that assuming a uniform prior in water:wine ratio is not the same as assuming a uniform prior in wine:water ratio, just as assuming a uniform prior in ECS is not the same as a uniform prior in λ , or $1/ECS$. The importance of this point depends on the level of uncertainty: the 5-95% range on ECS is still relatively large, making the impact of the prior distribution quite dramatic.

The resolution of Bertrand's paradox is to clarify the question: *why* are we interested in knowing the

ratio of water to wine? If we have drunk a glass of the mixture and want to know the amount of wine we have drunk, then the answer is somewhere between 0.33 and 0.5, with a best guess of 0.42, corresponding to a ratio of 7:5. The key is to apply the principle of indifference (uniform prior) to the quantity we are interested in, not to incidental quantities that are used to derive it, like the water-to-wine ratio or the ECS.

The only circumstance in which ECS is directly relevant to policy is when we are considering the response to an indefinite stabilisation scenario and quantifying the risk of climate impacts that are linear in the long-term equilibrium warming. A strict interpretation of Article 2 of the UNFCCC suggests that this is the principal objective of policy, in which case the relevant prior to use is uniform in ECS. Unfortunately, efforts to provide a range of uncertainty in ECS beginning from this prior, whether constraining ECS through the analysis of atmospheric feedbacks, through the comparison of climate models with climatology or through analysis of the transient response all continue to yield a high upper bound. The reason is that the first two approaches tend to provide information on λ , while the third approach yields information on TCR, neither of which is linearly related to ECS, with $d(\lambda)/d(ECS)$ and $d(TCR)/d(ECS)$ (which determine their information content with respect to sensitivity) both tending to zero as ECS increases. If the 95th percentile on ECS exceeds 7K, for example, then there is a significant (>5%) risk of a warming in excess of 4K even if atmospheric concentrations are stabilised at 450ppmv. Hence, if we assume an indefinite stabilisation scenario, there is a sense in which no stabilisation target currently under discussion can be considered “safe”.

How realistic, however, is an indefinite stabilisation scenario? Because of the relationship between sensitivity and adjustment time, these high-end sensitivities only materialise if concentrations are held constant at the stabilisation target for many decades, or even centuries. It is surely more plausible that future generations, having already reduced greenhouse gas emissions by 60-80% to achieve stabilisation, will be able to continue reducing emissions sufficiently to allow atmospheric concentrations to decline. Even a modest rate of decline (less than 10% over the first 40 years of the 22nd century after following the B1 scenario to 2100) means that the peak warming ceases to scale with ECS (figure d). In this simple model, the peak warming under such a concentration path is close to linear in both TCR and past attributable warming (figure e). TCR is also a better predictor than ECS for transient warming to 2100 under both A1FI and B1 scenarios [5,6] and for the warming commitment due to past emissions (the peak warming if emissions were to cease tomorrow).

Hence we propose that studies attempting to constrain climate sensitivity objectively will be most relevant to many policy questions if they begin by applying the principle of indifference to (assuming a uniform prior in) TCR rather than ECS or λ . This is almost exactly equivalent to assuming a uniform prior in past attributable warming [9]: the posterior distribution for ECS assuming this prior is shown by the black curve in figure (c). In this simple comparison with the transient response over the 20th century, this yields a 10-90% range for climate sensitivity of 1.3-4.5K, with a median value of 2.5K, in remarkably close agreement with the “classic” IPCC range of uncertainty [10].

References

1. N.G. Andronova and M.E. Schlesinger. Causes of global temperature changes during the 19th and 20th centuries. *Geophysical Research Letters*, 27:2137–3140, 2000.
2. C. E. Forest, P. H. Stone, A. P. Sokolov, M. R. Allen, and M. D. Webster. Quantifying uncertainties in climate system properties with the use of recent climate observations. *Science*, 295:113–117, 2002.
3. R. Knutti, T.F. Stocker, F. Joos, and G.K. Plattner GK. Constraints on radiative forcing and future climate change from observations and climate model ensembles. *Nature*, 416:719–723, 2002.
4. S. Levitus, J. Antonov, T. Boyer & C. Stephens, Warming of the world ocean. *Science* **287**, 2225–2229, 2000.
5. J. Hansen, G. Russell, A. Lacis, I. Fung & D. Rind, Climate response times: Dependence on climate sensitivity and ocean mixing. *Science* 229, 857–859, 1985.
6. M. R. Allen, P. A. Stott, J. F. B. Mitchell, R. Schnur and T. Delworth, Uncertainty in forecasts of anthropogenic climate change, *Nature*, **417**, 617-620, 2000.
7. P. A. Stott and J. A. Kettleborough. Origins and estimates of uncertainty in predictions of twenty-first century temperature rise. *Nature*, 416:723–726, 2002.
8. J. M. Murphy, D. M. H. Sexton, D. N. Barnett, G. S. Jones, M. J. Webb, M. Collins and D. A. Stainforth, Quantification of modelling uncertainties in a large ensemble of climate change simulations, *Nature*, to appear, 2004
9. J. M. Gregory, R. Stouffer, S. Raper, N. Rayner, and P. A. Stott. An observationally-based estimate of the climate sensitivity. *Journal of Climate*, 15, 2002. 3117-3121, 2002.
10. T. Wigley & S. Raper, Interpretation of high projections for global-mean warming. *Science*, **293**, 451-454, 2001.

Annex 2: Extended Abstracts

State of Radiation Codes in GCMs

Qiang Fu

Department of Atmospheric Sciences, University of Washington, Seattle, WA 98195, USA

The state of radiation codes used in current global climate models will be reviewed to highlight both main radiative transfer processes and current issues in modeling atmospheric radiation. Some basic aspects of radiative transfer processes including water vapor continuum, light scattering by nonspherical ice particles,

and effects of 3D radiative transfer will be emphasized. The main challenges as well as prospective approaches to complete our understanding of these processes will be addressed in terms of fundamental theory, modeling, and observations.

Annex 2: Extended Abstracts

Controls on the Interannual Variability of Planetary Albedo and Relevance for Climate Sensitivity

Alex Hall and Xin Qu

UCLA Dep't of Atmospheric and Oceanic Sciences

We assess the controls on planetary albedo variability by examining the ISCCP (Rossow and Garder, 1993a and 1993b) D2 data set, covering the period 1983-2000. For both clear and all-sky cases, the ISCCP data set (D2) contains surface and top-of-the-atmosphere radiation fluxes on a 2.5° X 2.5° grid at 3 hour temporal resolution. These were generated based on observations at 3 different channels (visible, near IR, and IR) and a radiative transfer model. The current D2 data set contains improvements over previous ISCCP releases. These increase low-level cloud detection significantly and reduce biases in cloud optical thickness over snow and ice (Rossow and Schiffer, 1999).

Planetary albedo fluctuations may originate at the surface or in the atmosphere. Within the atmosphere, clouds are likely the main sources of planetary albedo fluctuations because their high reflectivities create a large albedo contrast between a clear and cloudy atmosphere. For this reason, we focus on separating contributions from surface and cloud. Seasonal-mean planetary albedo anomalies are regressed onto seasonal-mean surface albedo anomalies, seasonal-mean cloud anomalies associated with cloud cover variations and seasonal-mean cloud anomalies associated with cloud optical thickness variations.

Based on this regression analysis, the variance of planetary albedo (left hand side of the following equation) can be attributed to four terms:

$$\langle a_p' \rangle^2 = \langle a_{ps}' \rangle^2 + \langle a_{pc}' \rangle^2 + \langle a_r' \rangle^2 + \langle a_{ps}', a_{pc}' \rangle^2 \quad (1)$$

where I. is the planetary albedo variance unambiguously related in linear fashion to surface albedo variability, II. is the variance unambiguously related in linear fashion to cloud cover and optical depth variability, III. is the residual component that cannot be linearly related to either surface or cloud variability, and the covariance term IV. is the portion linearly related to surface and cloud variability but not unambiguously attributable to either because of correlations between surface optical properties and cloud.

We divided the earth into six regions and analyzed them separately. Our choice of regions is guided by known differences in the behavior of surface albedo

variability: (a) northern hemisphere snow-covered lands, (b) northern hemisphere sea ice zone, (c) southern hemisphere sea ice zone, (d) snow-free lands, (e) ice-free ocean, and (f) Antarctica. Note that the definition of the regions varies seasonally. We averaged the contributions of the four components in Eq. (1) over each region for each season after first normalizing by the total planetary albedo variability (see figure 1). We will refer to this figure to compare the contributions of surface and clouds to planetary albedo variability among regions and among seasons within the same region.

An examination of the light grey bars of Fig 1 reveals that the contribution of the residual (component III in Eq. (1)) is negligible compared to the total contribution of surface albedo, cloud cover and cloud optical thickness during most seasons in nearly all regions, implying that these are the factors contributing most to planetary albedo variability. This is also viewed as a validation of our linear regression analysis technique. The contribution of the covariance term (component IV of Eq. (1), white bars of Fig 1) is also generally small (less than 10%), though it may be non-negligible in the SH sea ice zone during MAM. A weak negative correlation between fluctuations in surface albedo and those in cloud cover and cloud optical thickness occurs in the SH sea ice zone at this time of year, possibly attributable to cloud-cryosphere interaction. There also appears to be a weak but consistently positive correlation between surface and cloud properties in the snow-free lands during all seasons, possibly due to cloud-vegetation interaction.

A comparison of the black and dark grey bars of Fig 1 shows that over the snowfree lands and ice-free oceans, the cloud contribution overwhelms the surface contribution during all seasons; however, the surface makes the dominant contribution to planetary albedo variability in snow and ice regions at nearly all times of year. The surface contribution is so much larger in the snow and ice areas mainly because of the large surface albedo variability associated with snow and ice fluctuations (not shown). The surface contribution is also larger in the SH sea ice zone than its NH counterpart in all seasons. This is likely because surface albedo varies more in the SH sea ice zone at all times of year. In Antarctica, the surface contribution overwhelms the

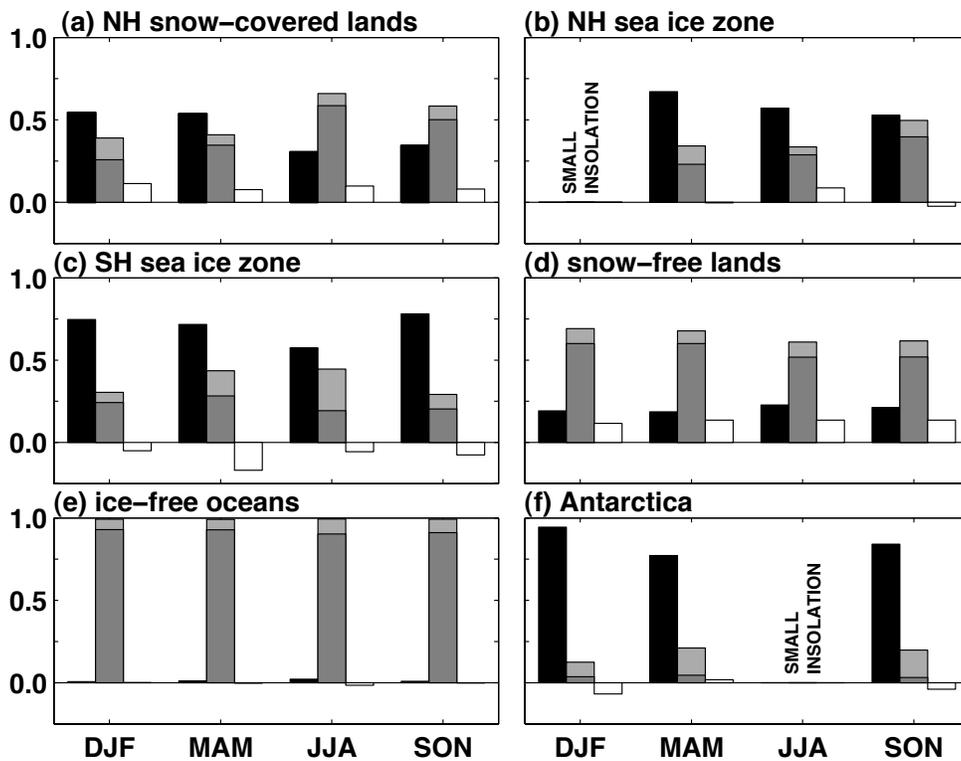


Figure 1. The contributions of surface albedo variations (black bar), cloud variations (dark grey bar), the covariance between surface albedo variations and cloud variations (white bar) and the residual (light grey bar) to planetary albedo variability averaged over six regions: (a) NH snow-covered lands; (b) NH sea ice zone; (c) SH sea ice zone; (d) snow-free lands; (e) ice-free oceans and (f) Antarctica. Terms I-IV of Eq. (1) were divided by the variance in planetary albedo (left side of Eq. (1)) to generate a normalized contribution to planetary albedo variability from each constituent. These were then averaged over the six regions.

cloud contribution during all seasons with sunshine. This can largely be explained by two facts: One is that the relatively transparent SH polar atmosphere does little to attenuate the impact of surface albedo anomalies on top-of-the-atmosphere radiation; the other is that cloud-related planetary albedo variability is negligible in this region.

The dominance of the surface contribution to interannual planetary albedo variability in the cryosphere regions and the apparent inability of clouds to mask the signature of surface albedo anomalies in net incoming solar radiation in these areas supports the positive surface albedo feedback seen in many future climate simulations. Just as present-day anomalies of surface albedo owing to variations in the cryosphere result in anomalies in net incoming solar radiation, an increase in net incoming solar radiation would occur if snow and ice were to retreat in the future.

The results presented in figure 1 can also be used to evaluate surface albedo processes in current climate models. To allow for as direct a comparison with the ISCCP data as possible, we used a simulated time series with approximately the same mix of internal variability and externally-forced climate change. A recent CCSM3

scenario run was used, with data taken from the same time period as ISCCP (1983-2000). We performed analysis identical to that shown in figure 1 on this data, and present the results in figure 2. A comparison of these two figures reveals that in all cryosphere regions, the contribution of the surface to planetary albedo variability is significantly smaller in CCSM3. It turns out this is because CCSM3 has substantially less surface albedo variability (not shown) than ISCCP in snow and ice regions. Understanding the reasons for this discrepancy may lead to ways to improve CCSM3's surface albedo parameterization, and in the process, improve its simulation of surface albedo feedback.

References

- Rossow, W.B., and L.C. Garder, 1993a: Cloud detection using satellite measurements of infrared and visible radiances for ISCCP. *J. Climate*, 6, 2341-2369.
- Rossow, W.B., and L.C. Garder, 1993b: Validation of ISCCP cloud detections. *J. Climate*, 6, 2370-2393.
- Rossow, W.B., and R.A. Schiffer 1999. Advances in understanding clouds from ISCCP. *Bull. Amer. Meteor. Soc.* 80, 2261-2288.

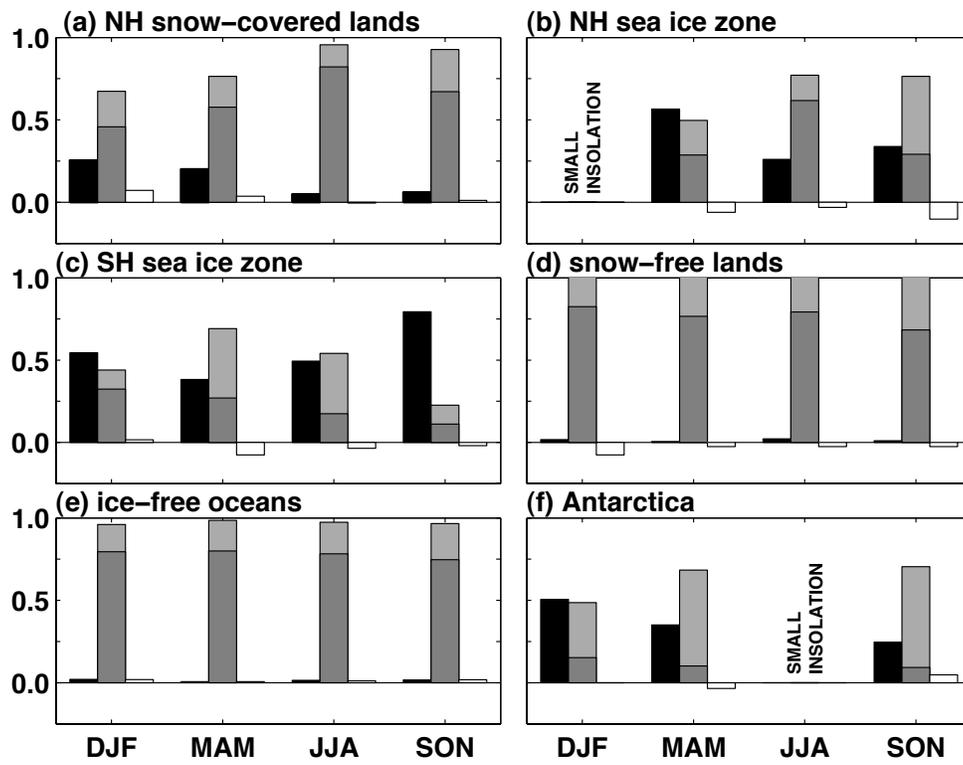


Figure 2. As in figure 1, except for the CCSM3 scenario run. Data was taken from years 1983-2000 of the run.

Annex 2: Extended Abstracts

Estimating Temperature Response to Radiative Forcing from Paleoclimate Records of the Last Millennium

Gabriele C. Hegerl and Thomas J. Crowley

Division of Earth and Ocean Sciences, Nicholas School for the Environment and Earth Sciences, Duke University, Durham, NC 27708

Paleoclimatic reconstructions of hemispheric temperature over the last few hundred to thousand years contain valuable information on climate response to external forcing such as volcanism, changes in solar radiation and increases in greenhouse gases. While the information on forcing and temperature response is less certain than during the instrumental period, the long records provide a better opportunity to separate the response to different external forcings that are spuriously correlated over the short instrumental period. Also, results based on paleoclimatic data can be used to validate our understanding of forced climate signals from the instrumental period. We have therefore used information in these reconstructions to detect temperature response to natural and anthropogenic forcings (Hegerl et al., 2003) and in an attempt to estimate climate sensitivity from paleoclimatic data (Hegerl et al., to be submitted shortly).

Paleoclimatic data

We have used temperature reconstructions based on tree-ring data that preserve low-frequency variability from Briffa et al. (2001) and Esper et al. (2002), an updated record from Crowley (2000) and the Mann et al. (1999) reconstruction to represent the uncertainty in our knowledge of past temperature fluctuations. Two of the records, Esper et al (2002) and Crowley (2000) are standardized averages of variations in paleoclimatic indices related to temperature. We have taken care to calibrate these to hemispheric scale temperature variability in a manner that preserves the magnitude of the radiative forcing in these records. This can be done by a regression approach that makes realistic assumptions about noise and uncertainty in both instrumental data and paleoclimatic data (see Allen and Stott, 2003). The resulting records of paleoclimatic variability show slightly more variability than in previously used calibrations. A perfect model test applying the same calibration procedure to synthetic data (using underlying climate signals from an Energy Balance Model and adding realistic amounts of noise to both, then use decadal data from 1880 to 1960 to calibrate) showed that the result yields reliable, robust calibration results.

Radiative forcing data

Long data for CO₂ forcing can be derived from ice cores. The estimate of the solar forcing time series is based on the Lean et al. (2001) solar irradiance reconstruction. It has been extended back by splicing the C14 residual record of Stuiver and Quay into it during the period of overlap. An updated version of a global volcano reconstruction (Crowley 2000) has been used to estimate past volcanic forcing (see Hegerl et al., 2003; Crowley et al., 2003). The time series utilizes only long ice core records of volcanism that extend back to A.D. 1200, namely four ice cores from Greenland and five from Antarctica produce a composite of sulphate flux for each hemisphere. A mid-line sulfate aerosol forcing is used from 1850 on. The temperature response to these external influences is simulated by a two-dimensional Energy Balance Model (EBM) with realistic geography and seasonal cycle. Care is taken to compare records with data from EBM simulations that represent the same seasonal (growing season or annual mean) and geographical (land only or land and ocean, latitude strips) representation as used in the paleoclimatic reconstruction.

We have first attempted to detect the temperature response to the different forcings by a multiple regression of the reconstructions onto the EBM fingerprints for volcanic forcing, greenhouse gas forcing (with superimposed aerosol forcing over the industrial period) and solar forcing. These fingerprints have been derived by EBM simulations forced with the individual forcing agents. Figure 1 shows the result using the Crowley reconstruction, results using the other reconstructions were comparable. The response to volcanic forcing is reliably detected in all reconstructions, and the simulated temperature response to volcanic eruptions compares favorably with observations. The response to solar forcing is detected in hemispheric mean data only over some periods in some records (those that represent annual rather than growing season data), and appears weak. Although most paleoclimatic records can be used only to the middle of the 20th century, the temperature response to CO₂ can be detected by this time in most records. Since the detection of the volcanic signal is dependent on a

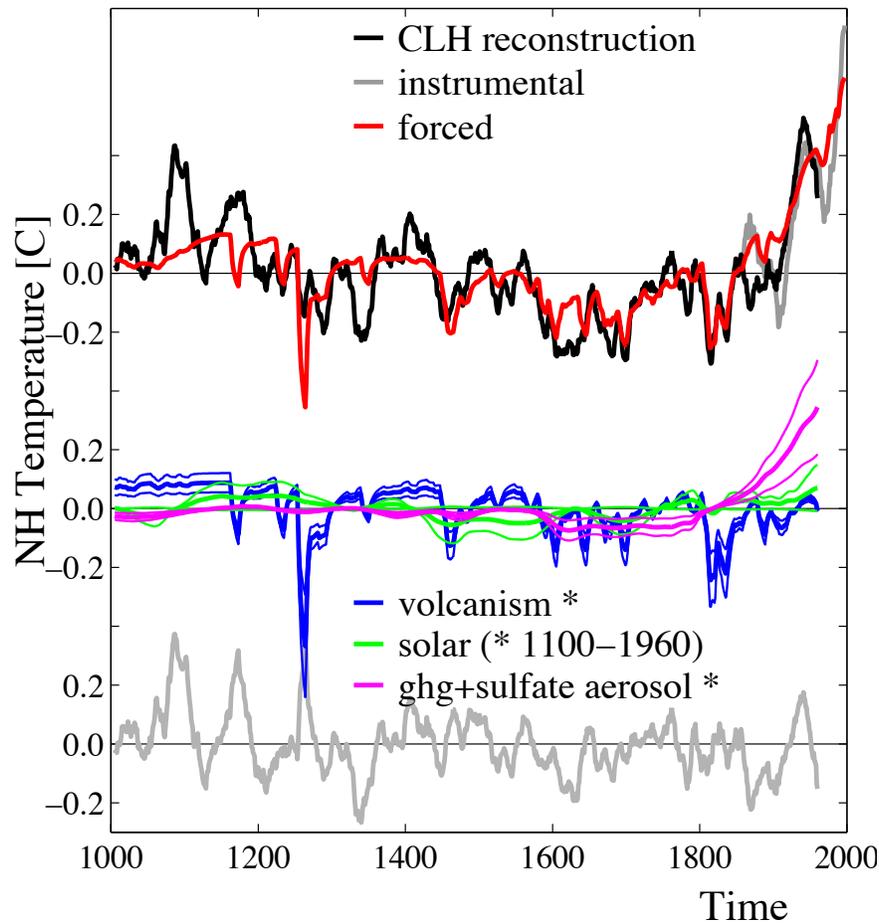


Figure 1. Detection results for the updated Crowley and Lowery (2000) reconstruction of decadal Northern Hemispheric mean temperature (north of 30N, calendar year average; from Hegerl et al., 2003). Upper panel: Paleo reconstruction (black) compared to the instrumental data (grey) and the best estimate of the combined forced response (red), middle panel: response attributed to individual forcings (thick lines) and their 5-95% uncertainty range (thin lines), lower panel: residual variability attributed to internal climate variability and errors in reconstruction and forced response. An asterisk "*" denotes a response that is detected at the 5% significance level. Results from other reconstructions are similar.

realistic representation of the temperature response to volcanism, it is important that particularly the response time to volcanism is realistic in the model. Figure 2 shows that this is the case.

Since the results from the detection of individual signals are encouraging, we have also used simulations forced with a combination of all relevant forcing agents in an attempt to estimate the probability density function (pdf) of climate sensitivity from paleoclimatic data. Our estimate of the pdf includes uncertainty in the paleo reconstructions, in solar and volcanic forcing, in ocean diffusivity, and in aerosol forcing. Results are quite comparable to those from instrumental data (Andronova and Schlesinger, 2000; Forest et al., 2000; Gregory et al., 2001) and will be shown at the workshop.

References

Allen M. R. and Stott P.A, *Climate Dynamics*, 21, 477-491 (2003).

- Andronova, N. G. and M. E. Schlesinger, *Geophys. Res. Lett.*, 27 (14), 2137-2140 (2000).
 Briffa, K. R., et al., *J. Geophys. Res.*, 106, 2929-2941 (2001).
 Crowley, T.J. *Science*, 289, 270-277 (2000).
 Crowley, T.J., et al., *Geophys. Res. Lett.* 30, doi:10.1029/2003GL017801 (2003).
 Esper, J., E. R. Cook, F. Schweingruber, *Science* 295, 2250 (2002).
 Forest, C.E., M.R. Allen, P.H. Stone, and A.P. Sokolov, *Geophys. Res. Lett.* 27, 569-572 (2000).
 Gregory, J. M. et al. *J. Climate* 15, 3117-3121 (2002).
 Hegerl, G. C. et al., *Geophys. Res. Lett.* 30(5), 1242. doi: 10.1029/2002GL016635 (2003)
 M. E. Mann, R.S. Bradley, M. K. Hughes, *Geophys. Res. Lett.* 26, 759 (1999).

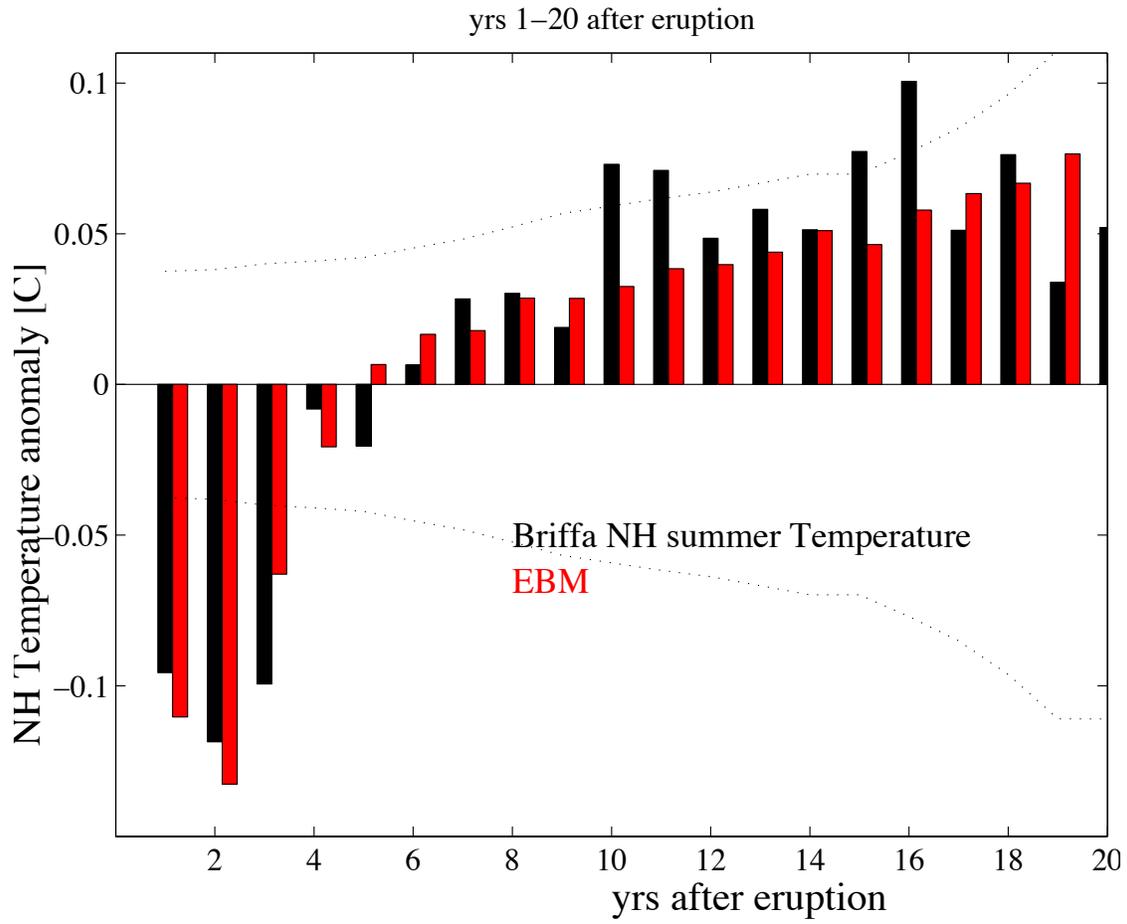


Figure 2. Comparison of the average response to volcanic eruptions in the energy balance model and the Briffa et al. (2001) reconstruction from the year of the eruption (year 1) to the next major eruption. 5-95% uncertainty ranges for the observed response are given by the dotted lines (note that sample size decreases with time). From Hegerl et al. (2003).

What Can Be Learned from Variations in Atmospheric CO₂ and Temperature over the Past Millennium?

Fortunat Joos

Climate and Environmental Physics, Physics Institute, University of Bern, Switzerland (joos@climate.unibe.ch)

Uncertainties in climate projections arise from uncertainties in natural and anthropogenic forcings and from the feedbacks between climate change and biogeochemical cycles, in particular climate-carbon cycle feedbacks. Here, the proxy-records of variations in NH temperature and in atmospheric CO₂ are applied within a model framework to constrain solar forcing, the contribution of solar forcing and solar forcing amplification to the 20th century warming, and to constrain the response of atmospheric CO₂ to modest climate change (Gerber et al., 2003; Ammann et al., 2004; Joos and Prentice, 2004). A probabilistic framework offers a way to assess uncertainties in climate projections in a self-consistent way and consistent with available observations while taking into account not only uncertainties in the climate sensitivity but also uncertainties in radiative forcing, in the carbon climate-cycle, and other uncertainties (Knutti et al., 2003).

Constraints on solar forcing and solar-induced climate change

Understanding and quantifying natural climate variability on decadal to centennial time scales is a prerequisite to detect and attribute anthropogenic warming and to project future climate change. It is important to extend the evaluation of models used for climate projections through the pre-industrial period when natural variations were pronounced while anthropogenic influence was small.

The magnitude of low-frequency solar irradiance changes is highly uncertain. Tentative correlations with records of cosmogenic nuclei (¹⁰Be, ¹⁴C), sunspots, aurora histories in combination with the behaviour of solar-like stars have been used to estimate past solar irradiance. The temporal evolution of different proxy series is in reasonable agreement for the past millennium. However, the scaling required to translate a proxy record into solar irradiance anomalies is highly uncertain and published estimates of multi-decadal solar irradiance changes vary by a factor of five. Apart from changes in total solar energy output, amplification of solar forcing and changes in stratospheric ozone distribution and wave dynamics or modifications of cloud properties by variations in cosmic-radiation are under discussion.

Various Northern Hemisphere (NH), and one global, surface temperature reconstructions for the past millennium have become available. These proxy-based reconstructions are affected by uncertainties and individual studies deviate notably from each other. However, taken together they suggest that natural low-frequency NH-temperature variations over the past millennium were within 0.3 to 0.9°C. The instrumental-based temperature record shows an increase in global average surface temperature of $0.6 \pm 0.2^\circ\text{C}$ over the 20th century.

Low-frequency solar irradiance changes were varied by a factor of 6.5 between different simulations with the NCAR coupled climate system model (Ammann et al., 2004). Simulations extended over the past 1150 years and include volcanic and anthropogenic forcing. Three key findings emerge.

First, the model is able to reproduce main features of the paleo temperature records. A clear link between the radioisotope-based solar irradiance record, NH-temperature reconstructions, and modelled temperature variations is found.

Second, large, low frequency solar irradiance variations are not compatible with the NH-temperature proxy records within the framework of the NCAR model. The range of model results encompass the range of reconstructed preindustrial NH-temperature variations. Given the low climate sensitivity of the NCAR model, smaller, possibly much smaller, than larger background trends in solar irradiance produce modelled climates in better agreement with the temperature proxy records.

Third, the cosmogenic isotope records and the NH-temperature proxy records constrain the contribution of natural climate variations to 20th century warming to be less than 0.2°C. All simulations with anthropogenic forcing included match the observed temperature increase over the industrial period and simulated global average surface temperature was higher during the most recent decades than during the previous 1100 years. On the other hand, only small warming over the industrial period is found in runs with solar and volcanic forcing only. This is the case even for the high solar scaling that yields larger than reconstructed temperature variations before the industrialization.

Constraints on climate-carbon cycle feedbacks.

Uncertainties in the response of the carbon cycle to future warming leads to uncertainties in projected warming of several degree Celsius. This is the same order of magnitude as uncertainties associated with the response of the physical climate system to a prescribed forcing, e.g. by greenhouse gases. Different carbon cycle – climate models have yielded different results on how future climate change affects the evolution of atmospheric CO₂. Projected increase in atmospheric CO₂ was found to be higher by a factor of 1.1 to 2 in simulations with global warming than in simulations without global warming. It is therefore imperative to consider the climate-carbon cycle feedback as well as the associated uncertainty in climate projections.

The term climate sensitivity is often defined as the response of the climate system to a unit change in radiative forcing and is expressed as the change in global average surface temperature for a nominal doubling of CO₂. The climate sensitivity of a model is then derived by prescribing a change in atmospheric CO₂ and evaluate the models temperature response. The response of the tropospheric water cycle, including changes in cloud cover, tropospheric water content, or changes in surface albedo through snow cover changes are included in most such assessments of the climate sensitivity. In contrast, the response of biogeochemical cycles and in particular the response of atmospheric CO₂ to climate change is not considered in this classical radiative forcing-climate sensitivity framework. In the IPCC TAR, models used to project atmospheric CO₂ included the climate-carbon cycle feedback. However, atmospheric CO₂ was prescribed in the TAR climate projections. Uncertainties associated with the climate-carbon cycle feedback have not been considered and consistency between the projected climate and the prescribed CO₂ is not guaranteed for a given emission scenario.

A probabilistic framework is useful to estimate the uncertainties associated with the climate-carbon cycle feedback, with forcings by various agents, and with the classical climate sensitivity. Uncertainties in natural and anthropogenic forcings both during the industrial and earlier period, combined with uncertainties in observations, and internal climate variability hamper our ability to constrain the climate sensitivity. Recent attempts to constrain the climate sensitivity from observations have shown that there exists the possibility that the change in global average surface temperature for a nominal doubling of CO₂ may well exceed the widely cited range of 1.5 to 4.5°C. The risk for high warming is further amplified by a potentially large positive climate-carbon cycle feedback. Higher temperature leads to

higher growth rate of atmospheric CO₂ and, in turn, to an amplified warming.

The ice core atmospheric CO₂ record combined with the range of NH temperature reconstructions covering the past millennium provides an opportunity to evaluate the climate-carbon cycle feedback. While the temperature variations over the past millennium were relatively modest compared to the projected changes, the proxy records form nevertheless an important yard stick for coupled carbon cycle-climate model.

Figure 1 illustrates the relationship between the variations of decadal to multi-decadal northern-hemisphere surface temperature, atmospheric CO₂ concentration, and the strength of the climate-carbon cycle feedback (as ppm CO₂ released per K temperature increase in the Northern Hemisphere). The strength of the feedback depends on several factors including the change in solubility of CO₂ in seawater and the responses of productivity and heterotrophic respiration to temperature and soil water. We assume that the pattern of climate change (temperature, precipitation, cloud cover), as far as relevant for atmospheric CO₂, remained approximately constant. The current best estimate of the actual CO₂ range during the past millennium (prior to the Industrial Revolution) is 6 ppm, based on emerging

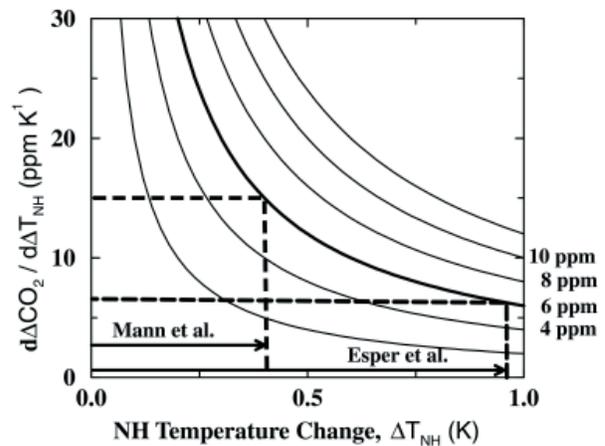


Figure 1. This diagram illustrates the relationship between northern-hemisphere (NH) surface temperature change, climate-carbon cycle feedbacks, and variations in atmospheric CO₂. Isolines depict different ranges for CO₂ variation during the last millennium and are plotted against changes in decadal-average NH-temperature (horizontal axis) and the climate-carbon cycle feedback expressed as change in atmospheric CO₂ concentration per degree change in decadal-average NH surface temperature (vertical axis). The range of NH-temperature variations reconstructed by Mann et al. (1999) and Esper et al. (2002) are shown by solid arrows. Combining the estimates for low-frequency NH-temperature variations and CO₂ variations yields a climate-carbon cycle feedback range of 6 to 16 ppm K⁻¹ (dashed).

high-quality measurements (Siegenthaler and Monnin, pers. comm.). If we accept the ranges of temperature variations reconstructed by Mann et al. (1999) and by Esper et al. (2002) as equally possible, then the CO₂ concentration range of 6 ppm constrains the climate-carbon cycle feedback to between 6 and 16 ppm K⁻¹ (for global mean surface temperature changes of less than ~1°C). This implies that the feedback of 12 ppm K⁻¹ found with the IPCC TAR version of the Bern carbon cycle-climate model in transient simulations over the past millennium is compatible with the proxy records. On the other hand, models with a very strong climate-carbon cycle feedback (for modest climate change) are not compatible with the CO₂ and NH-temperature proxy records.

Selected References

(pdf-files on <http://www.climate.unibe.ch/~joos/publications.html>)

Ammann C.M., F. Joos, D. S. Schimel, B. L. Otto-Bliesner, R. A. Thomas, Constraining the range of solar irradiance changes during the last millennium, in preparation.

Esper, J., E. R. Cook, and F. H. Schweingruber. Low-frequency signals in long tree-ring chronologies for

reconstructing past temperature variability. *Science*, 295: 2250-2253, 2002.

Gerber S., F. Joos, P. P. Brügger, T. F. Stocker, M. E. Mann, S. Sitch, and M. Scholze. Constraining temperature variations over the last millennium by comparing simulated and observed atmospheric CO₂. *Climate Dynamics*, 20, 281-299, doi: 10.1007/s00382-002-0270-8, 2003.

Joos F. and I. C. Prentice. A paleo-perspective on changes in atmospheric CO₂ and climate. In *The Global Carbon Cycle: Integrating Humans, Climate and the Natural World*, C. Field and M. Raupach (Eds.), SCOPE series 62, Island Press, Washington DC, USA, 165-186, 2004.

R. Knutti, T. F. Stocker, F. Joos, and G.-K. Plattner. Probabilistic climate change projections using neural networks. *Climate Dynamics*, 21, 257-272, 2003.

Mann, M. E., R. S. Bradley, and M. K. Hughes. Northern Hemisphere temperatures during the past millennium: Inferences, uncertainties, and limitations. *Geophysical Research Letters*, 26: 759-762, 1999.

Climate Sensitivity Inferred from Paleoclimatic Records

Jean Jouzel

Institut Pierre Simon Laplace / Laboratoire des Sciences du Climat et de l'Environnement CEA/CNRS CE Saclay 91191 Gif/ Yvette France

“Our most precise knowledge of climate sensitivity comes from data on ancient and recent climate change”. This sentence was not written by a paleoclimatologist but by J. Hansen and coworkers (1993) in an excellent review article “How sensitive is the world’s climate” published more than 10 years ago. These authors added that “the best empirical information on equilibrium climate sensitivity is provided by climate variations of the past 200,000 years”, the time span over which CO₂ and CH₄ data were then available from the Vostok Antarctic ice core, and derived a climate sensitivity $\Delta T_{2^*CO_2}$ of $3 \pm 1^\circ C$ for doubled CO₂.

As a paleoclimatologist, I was very pleased by this review article, but I am not so convinced that all climate modellers share Jim Hansen’s enthusiasm for paleoclimate estimates of climate sensitivity. Probably (e.g., Crowley, 1993), more has to be done to convince both paleoclimatologists and climatologists of the potentialities of paleodata in this respect, keeping them aware of the limitations inherent to this approach. This will hopefully be an outcome of our IPCC workshop which leaves ample room to this topic. Before these various presentations, it might be useful to briefly review published literature in this domain and this will be the main purpose of my talk. Very schematically, existing approaches fall in three categories with estimates respectively based on pre-Quaternary data (the distant past with both relatively badly known climate forcings and responses), on Quaternary data (indeed limited to the last few hundred thousands of years for which greenhouse gas data are available from ice cores), and on more recent data (the last millennia which is potentially interesting, even though the climate signal is weak). Among other examples, I will briefly examine the following with a focus on pre-Quaternary and Quaternary timescales.

Hoffert and Covey (1992) combined data from two paleoclimates, one colder (the Last Glacial Maximum, LGM, 20 000 years ago) and one warmer (the Mid-Cretaceous Maximum, MCM, about one hundred million years ago), to infer a $\Delta T_{2^*CO_2}$ of $2.1 \pm 0.4^\circ C$; see also criticism by R. Lindzen (1992) and Covey et al. (1996). However, it is clear that we have to be cautious going further back in time as recently exemplified by a paper by Shaviv and Veizer (2003). Starting from a reconstruction of cosmic ray fluxes and of its apparent

high correlation with temperature over the last 520 million years, these authors inferred that CO₂ has a smaller effect than previously thought with a $\Delta T_{2^*CO_2}$ as low as 0.75°C. Indeed, in Rahmstorf et al. (2004), we clearly point out that the claimed correlation between cosmic ray flux and temperature appears to not hold up under scrutiny and that in any case the author’s estimate of the effect of a CO₂ doubling is highly questionable.

The late Quaternary is probably more appropriate to provide estimates of $\Delta T_{2^*CO_2}$ although there are, here again, limitations. The advantages are that glacial-interglacial temperature changes are large (mean value of about 5°C) and that greenhouse gas changes are well documented as well as, albeit indirectly, other potential forcings such as the one linked with change in aerosol loadings. Although, the initial instigator of ice ages is the variation of insolation due to periodic changes of the Earth’s orbit, the contribution of greenhouse gases can be investigated. Prior to the above mentioned Hansen et al. estimate, we inferred from Vostok and other data (Lorius et al., 1990) climate sensitivity over the last 150 000 years (last climatic cycle). Over this time period the contribution of greenhouse gases was estimated to be between 40 and 65%—a figure however higher than inferred from climate model experiments as performed by Berger and colleagues (1993)—leading to the conclusion that a $\Delta T_{2^*CO_2}$ of 3 to 4°C may be a realistic value.

Extending this calculation to four climatic cycles (Vostok data), which has not yet been done, should not drastically change this estimate as the correlation between greenhouse gases and Antarctic climate does not vary significantly from one glacial cycle to the next. The situation might differ, and indeed appears quite promising, with the recent extension of the ice core record to 800,000 years (EPICA Dome C ice core). There is a change of pacing prior the last four climatic cycles with less warmer interglacial periods. This should offer, once greenhouse data will be available, the possibility to test situations (interglacials) with relatively similar boundary conditions and, hopefully, different greenhouse levels (see presentations of P. Braconnot and J. Overpeck dealing with interglacials). In addition one weakness of this approach, i.e. the fact that it deals with transient climate changes, should be, at least partly, overcome by the use of models of intermediate

complexity which allow to depict such transient climatic changes.

Obviously, the steady state approach followed by Hansen et al. (1993) and by many other groups using data from the Last Glacial Maximum, has also a lot to offer (see presentations of S. Rahmstorf and B. Otto-Bliesner). However they have limitations too, which also applies for transient estimates. Such limitations come, in particular, from the uncertainties in the estimate of the mean global surface temperature change and from the fact that climate sensitivity between the ice age and today may differ from that between today and a warmer world, due for example to cloud processes (e.g., Ramstein et al., 1998).

The use of data from the last millenium which will be discussed by G. Hegerl and N. Andranova, and dealing more specifically with the CO₂ cycle itself, by F. Joos, is more recent. On this timescale, I will focus on climate forcings, which are directly or indirectly estimated from ice core data, and on associated uncertainties.

References

- Berger, A., Loutre, M. F., Tricot, C. and Gallée, H., Water vapour, CO₂ and insolation over the last glacial-interglacial cycles, *Phil. Trans. R. Soc. Lond. B*, 341, 1993, p. 253-261.
- Covey, C., Sloan, L. C. and Hoffert, M., Paleoclimate data constraints on climate sensitivity : The paleocalibration method., *Climate Change*, 32, 1996, p. 165-184.
- Crowley, Geological assessment of the greenhouse effect, *Bulletin of the American Meteorological Society*, 74, 1993, p. 2363-2373.
- Hansen, J., Lacis, A., Ruedy, R., Sato, M. and Wilson, H., How sensitive is the world's climate, *Research and exploration*, 9, 1993, p. 143 - 158.
- Hoffert, M. and Covey, C., Deriving global climate sensitivity from paleoclimate reconstructions, *Nature*, 360, 1992, p. 573-576.
- Lindzen, R., Paleoclimate sensitivity, *Nature*, 363, 1993, p. 25-26.
- Lorius, C., Jouzel, J., Raynaud, D., Hansen, J. and Le Treut, H., Greenhouse warming, climate sensitivity and ice core data, *Nature*, 347, 1990, p. 139-145.
- Rahmstorf, S., Archer, D., Ebel, D. S., Eugster, O., Jouzel, J., Maraun, D., Neu, U., Schmidt, G., Severinghaus, J., Weaver, A. J. and Zachos, J., Cosmic Rays, Carbon dioxide and Climate, *E.O.S*, 85, 2003, p. 38, 41.
- Ramstein, G., Serafini-Le Treut, Y., Le Treut, H., Forichon, M. and Joussaume, S., Cloud processes associated with past and future climate changes, *Clim. Dyn.*, 14, 1998, p. 233-247.
- Shaviv, N. and Veizer, J., Celestial driver of Phanerozoic climate, *GSA Today*, 13, 2003, p. 4-10.

Climate Sensitivity - The Role of Aerosol Effects

Stefan Kinne

Max-Planck-Institute for Meteorology, Hamburg, Germany

INTRODUCTION

AEROSOL and CLIMATE

Industrialization and land-use changes have increased the aerosol load in the Earth's atmosphere. However, in what way this aerosol increase has influenced the available radiative energy to the Earth-Atmosphere-System (in particular the distribution of solar and terrestrial radiation) remains unclear. Uncertainties are related to the complexity in distribution and properties of aerosol and to interactions with other atmospheric particles (in particular clouds). In an effort to better address the different nature of these aerosol-climate complexities, it is commonly distinguished between the impact from the aerosol presence (*direct effect*) and the impact of aerosol induced modifications to other atmospheric properties (*indirect effects*).

FORCING

The climate impact for any change to the atmospheric composition is commonly quantified by the change of the radiative energy escaping to space. Increases to the energy loss stand for *cooling*. Reductions to the energy loss stand for *warming*. This so-called 'radiative forcing' at the top of the atmosphere (ToA) is (for more detail) often further separated by spectral region (solar and terrestrial) and by location (atmosphere and surface). To provide just one number, regional and seasonal forcing is commonly summarized by a globally and annual averaged forcing value. For anthropogenic aerosol such ToA forcing is believed to be a cooling. However, there is a strong modulation of this impact (even in sign) by region and season. One of the explanations is that aerosol forcing does not only depend on the various aerosol properties but also on environmental properties. Some of the critical parameters for aerosol forcing are introduced next.

DIRECT EFFECT

PARAMETERS

The direct ToA aerosol forcing is mainly modulated by aerosol properties (of concentration, size, absorption and altitude), by surface properties, by the presence of clouds and available sun-light.

Aerosol size determines spectral regions of importance. Aerosol covers several orders of

magnitude in size - from a few nanometers to at least ten micrometer. However, from a radiative transfer perspective only the larger aerosol sizes ($>0.1 \mu\text{m}$) are important. With a natural concentration minimum at $1.0 \mu\text{m}$ sizes (due to growth and removal processes) the larger aerosol particles are usually separated in two size modes. Aerosols of the smaller accumulation mode ($0.1 - 1.0 \mu\text{m}$) are largely 'anthropogenic', while aerosols of the larger coarse mode ($>1.0 \mu\text{m}$) are mainly of 'natural origin'. The smaller aerosol particles of the accumulation mode mainly influence the energy distribution in (the visible region of) the solar spectrum. The larger aerosol particles of the coarse mode affect the entire solar spectrum as well as part of the terrestrial spectrum.

Aerosol absorption at solar wavelengths has an important influence on the distribution of aerosol forcing within the atmosphere. Aerosol solar absorption (self-) heats its atmospheric layers. As this atmospheric forcing allows less solar (scattered) radiation to escape from the atmospheric aerosol layers, the forcing at the surface will be more negative, while at the same time aerosol (reflection associated) solar energy losses at the ToA are reduced (less cooling). In fact, ToA warming is expected from strongly absorbing aerosol.

Aerosol absorption at infrared wavelengths can trap terrestrial radiation that is otherwise lost to space. Such aerosol greenhouse ToA warming, however, will only be significant for the combination of elevated aerosol and larger particles of the coarse mode (e.g. dust). Many studies ignore the infrared component to the ToA aerosol forcing, because the altitude / large size combination is rare, and because even for those cases the solar (albedo) effect usually dominates the (infrared) greenhouse effect (on the basis of daily averages).

Solar surface albedo below the aerosol layer has a strong effect on strength and sign of (clear-sky) aerosol solar ToA forcing. Over dark surfaces (e.g. ocean) aerosol associated solar energy losses to space are maximized: ToA cooling can be expected – unless aerosol is strongly absorbing. Over bright surfaces (e.g. deserts and especially snow) aerosol additions to already high solar energy losses to space are small: ToA warming can be expected, because already weak aerosol absorption will lead to ToA warming.

Cloud presence complicates aerosol ToA forcing to the point that the relative altitude between aerosol and clouds must be known. If located below clouds

the aerosol impact becomes small, because the solar energy available for interactions with aerosol is strongly reduced. However, when aerosol is located above clouds, then these clouds act similar to bright surfaces with an overall tendency towards ToA warming. Aerosol associated ToA cooling under clear-sky conditions will represent an overestimate (on the order of 50%) with respect to all-sky conditions.

MEASUREMENTS?

Radiative forcing by nature is a difference to an aerosol (component) free or less loaded (e.g. pre-industrial) reference state. Thus, a measurement of the aerosol direct radiative forcing is basically impossible. At best, a combination of different satellite sensors may provide clear-sky forcing estimates (e.g. EOS Terra: MODIS/MISR → aerosol optical depth in cloud free regions; CERES → associated ToA fluxes). However, those estimates will be limited to few regions and seasons where the ToA solar aerosol signal exceeds statistical noise. In addition, measurements cannot distinguish between natural and anthropogenic contributions (although new efforts involving size information [assuming sub-micron sized aerosol to be of anthropogenic origin] may provide rough estimates). Thus, when interested in the impact on anthropogenic aerosol on climate (globally and under realistic all-sky conditions) we need to turn to modeling. And our current understanding on the influence of aerosol on climate is almost entirely based on model simulations.

MODELING

Coarse gridded (ca. 200*200km, 20 vertical levels) global circulation models (GCM) or chemical transport models (CTM) are applied to provide estimates for the aerosol impact on climate. In recent years almost all models added more detail to their aerosol modules to improve aerosol representation (concentration, size, hygroscopicity, absorption and altitude) and to improve aerosol processing (e.g. subscale variability, cloud schemes). In recognition, that aerosol has a lifetime of a few days and originates from a multitude of different sources, most aerosol modules in these global models now distinguish among at least five different aerosol components:

BC ('Black Carbon' or **EC** 'elementary carbon') originates from wildfires and consumption of bio-fuel and fossil-fuel. Black carbon is a strongly absorbing aerosol of accumulation mode. An initially hydrophobic behavior is moderated with time as mixtures with other components occur.

OC ('Organic Carbon' or **POM** [$\Leftrightarrow 1.4*OC$] 'Particulate Organic Matter'), originates (as BC but more abundant) from wildfires and the consumption of

bio-fuel and fossil-fuel. A minor fraction also enters via the gas phase. Organic carbon is a weak to moderately absorbing aerosol of the accumulation mode. Organic carbon is moderately hydrophilic (which means this aerosol type will increase in size as the ambient relative humidity increases).

SU ('sulfate') predominantly enters the atmosphere via the gas-phase from fossil fuel consumption, volcanic exhaust and oceanic phytoplankton (DMS – dependent on solar surface flux). Sulfate is a non-absorbing aerosol of the accumulation mode. Sulfate is hydrophilic.

DU (mineral 'dust') originates from mainly from dry lake beds. Critical parameters are near-surface winds, soil moisture, vegetation and snow cover. Dust is (excluding strong UV absorption) a weak absorbing aerosol of the coarse mode and usually considered hydrophobic.

SS ('sea-salt') aerosol originates from sea-spray. The most critical parameter is the near- surface wind. Sea-salt aerosol is a non-absorbing aerosol of the coarse mode. Sea-salt is highly hydrophilic and at high ambient humidity aerosol size is comparable to cloud droplet sizes.

Aerosol at any particular location and time is always a mixture of aerosol components. While some of the aerosol components are externally mixed, internal mixtures of components occur as well. Since optical or hygroscopic properties of internal mixtures usually differ from volume weight averages of individual components, some of the more advanced aerosol modules in global modeling now even consider internal mixing. Aerosol component modeling is usually done in three consecutive steps:

STEP 1: Based on (particle and gas) emission inventories for each aerosol type, global models process aerosol (e.g. gas-to-particle conversion, growth, removal, redistribution by clouds) to yield mass concentration.

STEP 2: Mass is converted into (the optical property of) aerosol optical depth (as a function of wavelength), where assumptions for aerosol size (-distribution) and water uptake (humidification and ambient relative humidity) are critical.

STEP 3: Aerosol optical depth spectral data along with spectral data on aerosol absorption (based on absorption assumption of contributing aerosol types) are processed in radiative transfer schemes to yield an aerosol forcing.

Each of these steps introduces uncertainties. Unclear in particular are uncertainties introduced in *STEP 1*, where in contrast to *STEP 2* (measurements supply data) or *STEP 3* (accurate methods exists) reference data are lacking. In addition, with the increased complexity of aerosol modules many new assumptions were introduced, which have not been completely tested. Thus, at least at this stage the

uncertainty in modeling actually may have increased. Even worse, our current understanding of uncertainty may be biased.

UNCERTAINTY

Uncertainty is often not what it appears to be. Differences for the simulated end-products (ToA radiative forcing) are not necessarily a good estimate for forcing uncertainty, if aerosol processing (e.g. emission strength, transport, removal, chemistry in *STEP 1*) is not completely understood and/or poorly evaluated. Currently most models to some degree ‘tune’ aerosol processing at a subsequent step to readily available data on aerosol optical depth (e.g. global fields from satellite remote sensing or data from local AERONET statistics). To emphasize this point, simulated global annual averages for the mid-visible aerosol optical depths among 15 global models agree within a factor of 2. More importantly however 10 models fall in the 0.126-0.145 range. For these 10 models maximum-minimum factors are displayed in Figure 1 on a component basis for optical depth, dry mass and mass extinction

efficiency (the ratio of the former two properties). These large differences raise many questions. Here are some:

QUESTIONS

- Are aerosol forcing simulations (via emission data) accurate enough to be useful?
- Is modeling still missing important processes and interactions (e.g chemistry, scale)?
- Are individual models sufficiently explored in terms of strength and weaknesses?
- How to place resources in modeling (detail overkill vs. detail need)?

INDIRECT EFFECTS

Aerosol indirect effects summarize the climatic impact of all aerosol related modifications to other atmospheric properties. Particular important are aerosol induced changes to clouds, because clouds are the main modulators to the distribution of radiative energy in the atmosphere. Aerosol can influence clouds in many ways. Some of these effects support ToA cooling (e.g. smaller droplets, extended lifetime) while others support

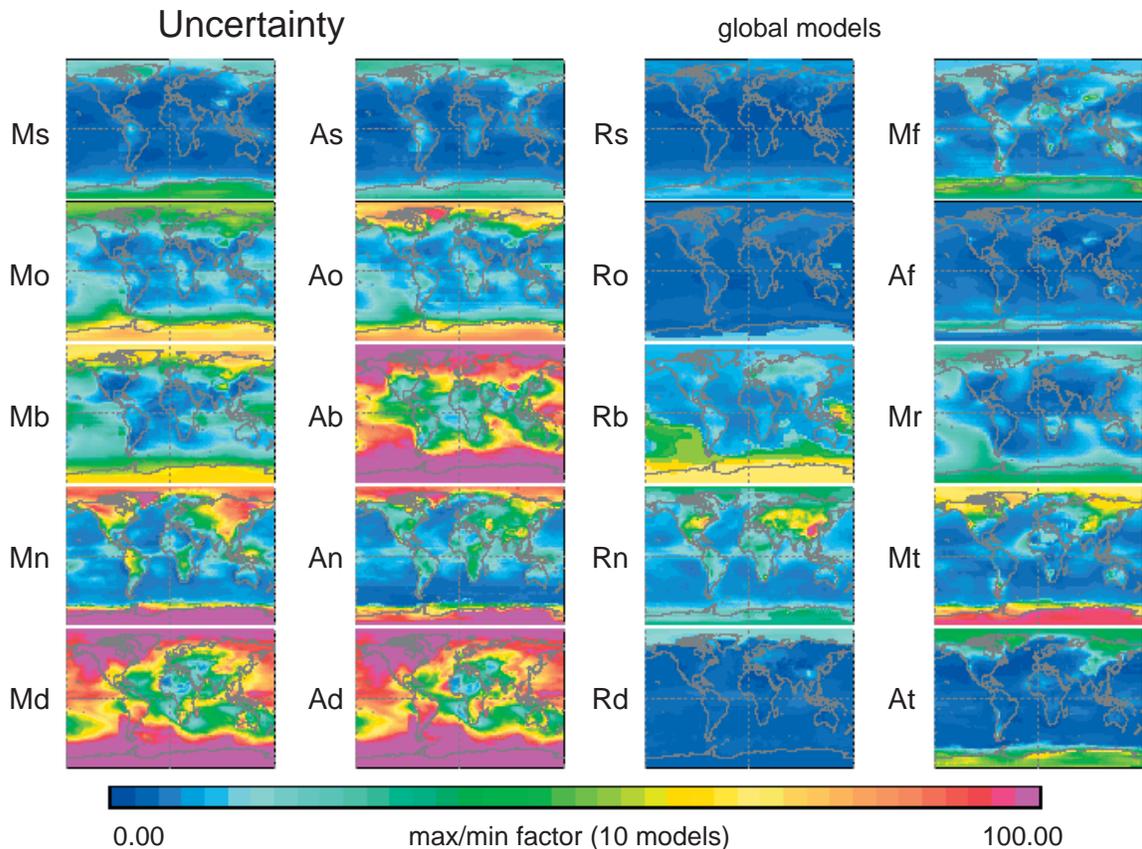


Figure 1. Uncertainty factors in global modeling based on local (1*1 degree) maximum / minimum ratios of (yearly averaged) monthly means of 10 global models (CTM or GCM). Aerosol properties for mass (M), mid-visible optical depth (A) and mass extinction efficiency (R) are addressed by component (s – sulfate, o – organic carbon, b – black carbon, n – seasalt, d – dust). The right row of panels presents totals (t) of all components, fractions (f) attributed to small sizes (<1µm) and the bc/oc mass ratio (r). Note, that all 10 global models agree to better than 20% for the annual, globally averaged mid-visible aerosol optical depth.

ToA warming (e.g. suppressed convection). Thus, even the sign of the overall indirect effect is not assured. Process studies (e.g. shiptracks) often provide only a distorted picture, because many indirect effects are often temporally delayed and locally detached (e.g. changes in precipitation patterns). Besides from properties of clouds and environment properties (e.g. background aerosol) also the aerosol properties are essential. In light of the large uncertainties regarding the aerosol direct effect several questions come to mind:

QUESTIONS

- Can we learn more by looking at observational data than relying on modeling?
- Are we aware of all aerosol indirect effect (and associated feedbacks)?
- Are cloud schemes in global model sufficiently accurate for aerosol - cloud interactions?

ACTIONS

DIRECT EFFECT

Different approaches are taken to improve uncertainties of the direct aerosol forcing. Two of those are outlined below. The first approach is a measurement-based diagnostic for aerosol modules. Its goal is to identify and eliminate weak modeling components, to quantify uncertainties at each modeling step and translate these uncertainties into a forcing uncertainty. The second approach seeks to apply data (and data uncertainty) directly into forcing, circumventing many model uncertainties associated with aerosol processing. Yet, even these more data oriented approaches require still significant model support.

1. To better quantify the skill of aerosol component modules in global modeling and to restrict modeling freedom, measurements are the key. This is recognized the AeroCom activity. Its goals are to diagnose (weaknesses in) aerosol modules and to quantify uncertainties at each modeling *STEP* and for each aerosol component (e.g. *SU*, *BC*, *OC*, *SS*, *DU*). Any proper assessment of differences between models with data requires an understanding for data quality (which is often worse than promised) and data representation. Evaluations are mainly based on available statistics of ground based in-situ sampling (e.g. IMPROVE network), of ground-based remote sensing (e.g. AERONET or EARLINET networks) and remote sensing from space (e.g. MODIS and MISR sensor data). AeroCom has initiated consistency tests for aerosol processing by requesting model-output for prescribed (yr 2000) harmonized input on emission and meteorology. Special attention will be given to outliers, as outliers indicate a deficiency, either by one or by all other models. Also, consistency among models is not necessary

a measure of uncertainty (e.g. models agree best for organic carbon aerosol, one of the least understood aerosol components). AeroCom's goal is to quantify 'real' uncertainty at sub-steps in forcing simulations and to translate these into overall forcing.

2. With new and more capable sensors from remote sensing and better statistics from ground data, there are now efforts to tie direct aerosol forcing estimates stronger to observational data. For instance, detailed data on aerosol properties from ground statistics (e.g. AERONET inversions provide data on concentrations, absorption and size). These data can be merged with satellite remote sensing of aerosol, clouds and surface (solar albedo). Aerosol direct forcing estimates from such data are presented in Figure 2. Other ideas involve better estimates based on the merging of data-streams from measurements and global modeling. For instance, (global) modeling can be applied to extend sparse sampling in space and time or models can provide needed information not available from measurements (e.g. forcing efficiency). Still, the fact that these methods are under considerations reflects the overall large uncertainty associated with aerosol direct forcing, especially on a regional basis.

INDIRECT EFFECTS

Also several approaches are taken improve uncertainties for the aerosol indirect forcing, by recognizing and eliminating poor assumptions. Two of those approaches are outlined below. The first approach is based on the exploration of correlations among retrieved aerosol and clouds from space observations. The second approach focuses on sensitivities to particular assumptions in aerosol-cloud interactions among models.

1. Correlations between 'quasi-simultaneous' retrievals of clouds and aerosol can be used to locate hot-spots for aerosol cloud interactions. Correlation maps serve as incentive to studies on important and dominant processes and as a footprint that needs to be matched by models (to demonstrate modeling skill). Under the AeroCom exercise correlations of aerosol and cloud properties are explored, based on daily retrievals with the MODIS sensor for the year 2000. Modelers are asked to simulate with (prescribed) year 2000 aerosol emissions and year 2000 meteorology to assure similar background conditions.

2. Under auspices of the US lead Climate Change Science program (CCSP) (and in close collaboration with AeroCom) a model-intercomparison of aerosol-cloud interactions in global modeling has been proposed. For several control experiments, with different degrees of freedom, the output of several models (including the associated forcing) is compared. Goal of this exercise is to identify and remove estimates of models with poor processing from uncertainty estimates.

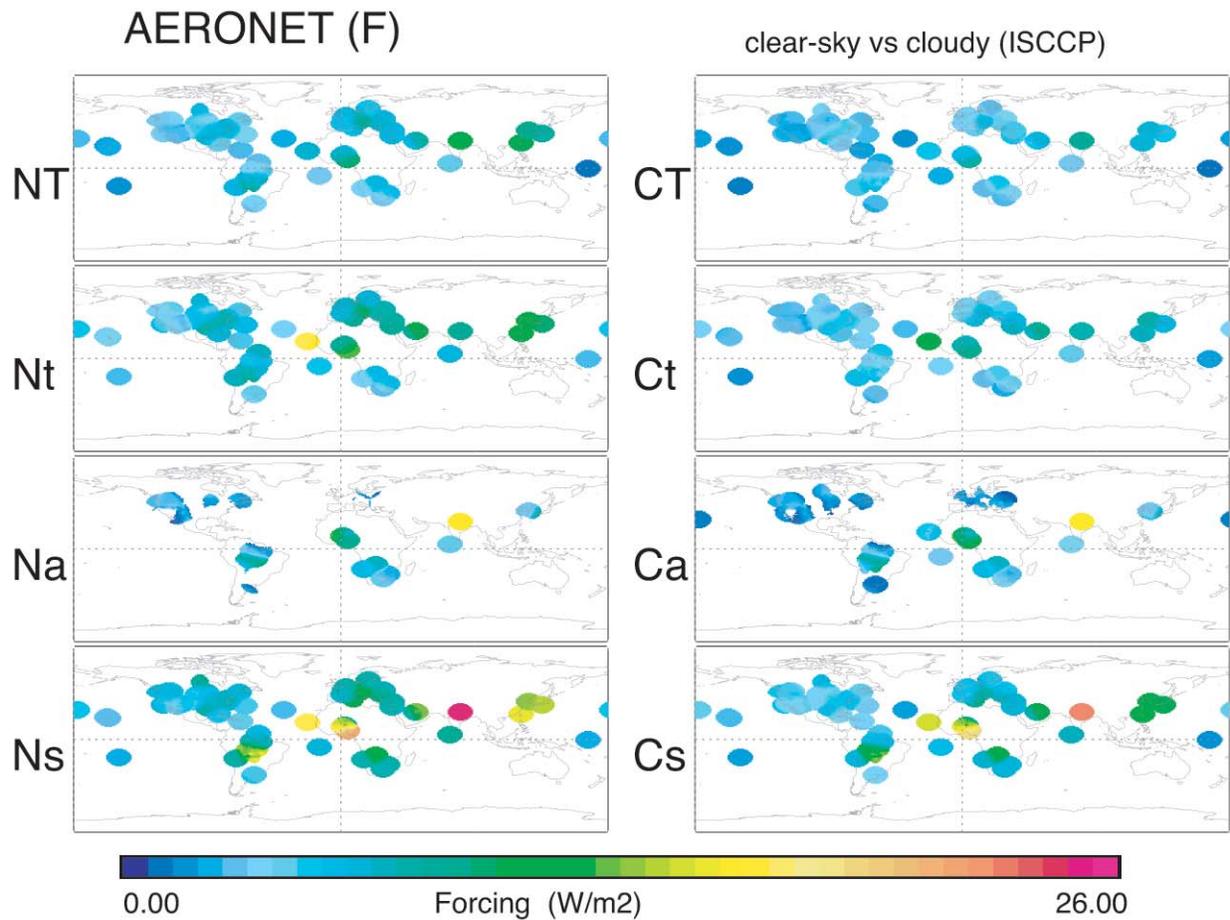


Figure 2. Aerosol Forcing based on 1998-2001 AERONET data. The presented yearly means are based on monthly statistics. Separate results are shown for forcings at the top of the atmosphere (t), in the atmosphere (a) and at the surface (s). The left column of panels shows the forcing under clear-sky condition (N), the right column of panels shows the forcing under cloudy conditions (using ISCCP cloud data). Also shown in the first row are forcings at the top of the atmosphere solely due to aerosol sizes of the accumulation mode (T, as anthropogenic estimate).

A Comparison of Climate Sensitivity Among the Japanese Models

Akio Kitoh¹, Akira Noda¹, Seiji Yukimoto¹, Seita Emori², Toru Nozawa², Tomoo Ogura², Masahide Kimoto³, Ayako Abe-Ouchi³, and Akimasa Sumi³

¹ Meteorological Research Institute, Tsukuba, Japan

² National Institute for Environmental Studies, Tsukuba, Japan

³ Center for Climate System Research, University of Tokyo, Tokyo, Japan

1. Introduction

Two Japanese models (MRI and CCSR/NIES models) showed the smallest and the largest climate sensitivity at the time of the IPCC TAR in 2001. Here we report recent changes in their models' climate sensitivity.

2. MRI model

The MRI-CGCM2 (Yukimoto et al. 2001; Yukimoto and Noda 2002) has several versions depending on cloud-radiation tuning. In its radiation scheme, CO₂, H₂O, O₃, CH₄ and N₂O are treated as the greenhouse effect gases directly, and the direct effect of sulfate aerosol is explicitly treated, but the indirect effect is not included. For the AR4 scenario experiments, atmospheric concentration of sulfate aerosols has been calculated for all SRES emission scenarios of sulfur dioxide by a chemical transport model MASINGER developed at MRI (Tanaka et al. 2003). The global mean surface air temperature rises about 2.4K, 2.7K and 2.0K at the end of the 21st century compared to the 1961-1990 value in A1B, A2 and B2 scenarios, respectively (Uchiyama et al. 2004).

The global and annual mean surface air temperature change at 2xCO₂ of the MRI-CGCM2.0 in the IPCC TAR was very small, that is, 1.1K, with effective

climate sensitivity of 1.4K. In this simulation, the energy budget was not balanced between the top and the bottom of atmosphere, and the global mean and meridional distribution of radiative flux at the top of atmosphere had a large difference from the observation. The revised MRI-CGCM2.3 is improved in reproducing mean climate of global-mean and meridional distribution of energy budget by mainly adjusting clouds representation. With the improved version of the model, the effective climate sensitivity increased more than 1K. The decrease of negative feedback due to cloud forcing can explain most of the changes in climate sensitivity. The major difference of cloud feedback can be explained by the change in tropical low-level clouds for shortwave forcing and in tropical middle level clouds for longwave forcing. Associated with the tropical low-level clouds, the stratus and stratocumulus clouds over the eastern part of the oceans (especially over the eastern Pacific) have large impact on shortwave forcing. The change in the tropical middle level clouds is associated with the change in deep convective clouds.

In summary, the low climate sensitivity in the TAR SRES simulation with the MRI-CGCM2.0 is mostly attributable to an insufficient tuning of the cloud-radiation process. Simulations with a revised version

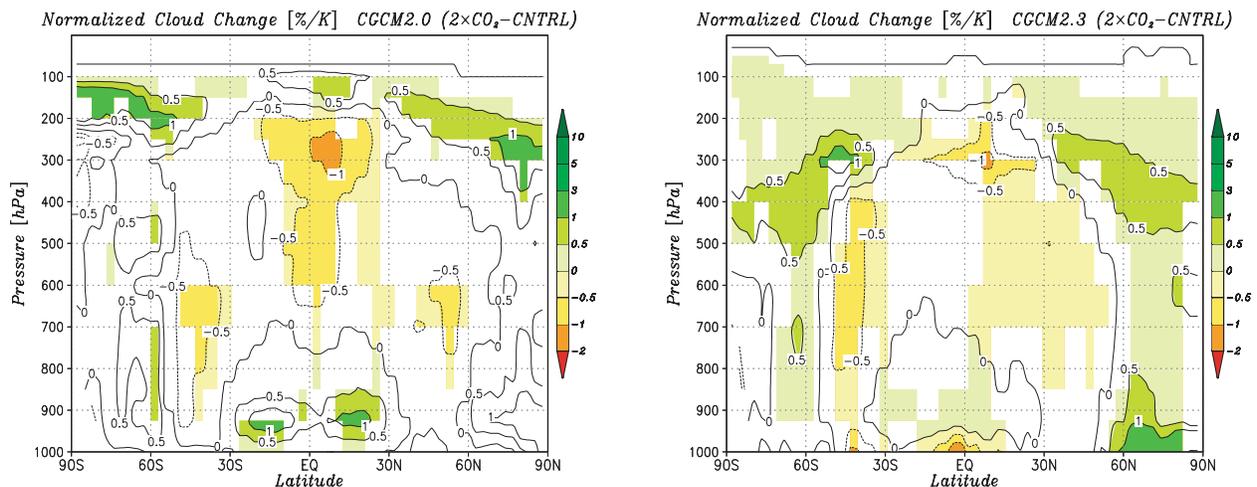


Figure 1. Zonally averaged annual mean cloud amount change normalized by global-mean surface air temperature change for (left) MRI-CGCM2.0 and (right) MRI-CGCM2.3.

Table 1. Climate sensitivity of the MRI-CGCM2.3 and two CCSR/NIES2 versions.

	MRI-CGCM2.3	CCSR/NIES-L	CCSR/NIES-H
Equilibrium sensitivity	3.20 K	3.95 K	6.30 K
Transient (yr61-80) climate response	2.18 K	2.11 K	not measured
Equilibrium precipitation change	7.33%	8.75%	13.95%
Transient precipitation change	4.49%	3.52%	not measured

of the MRI-CGCM2.3 show a good agreement in the global mean surface temperature observed in the 20th century.

3. CCSR/NIES/FRSGC model

The CCSR/NIES/FRSGC model (CCSR/NIES1: Emori et al. 1999) has been updated by including RIAM/CCSR aerosol transport-radiation model (SPRINTARS; Takemura et al. 2000), which can handle major tropospheric aerosols (carbonaceous, sulfate, sea salt and soil dust), and also has a parameterization of both the 1st and 2nd kind indirect effects of aerosols on cloud (albedo and lifetime). This updated model (CCSR/NIES2) is described by Nozawa et al. (2003).

The global mean surface air temperature rises about 4.5K, 3.3K, 5.3K and 3.9K at the end of the 21st century compared to the 1961-1990 value in A1B, B1, A2 and B2 scenarios, respectively. These values are higher than those in the CCSR/NIES1 due to a use of a different set of absorption parameters for various gases in the radiation code and due to the difference in cloud feedback (positive in CCSR/NIES2 and negative in CCSR/NIES1). About two-thirds of the difference is explained by the former (Nozawa et al. 2003).

The CCSR/NIES2 in the TAR shows the climate sensitivity of 5.1K. The reason of generally high sensitivity is that the model shows a decrease in low-level cloud cover by warming, thus having a positive shortwave cloud feedback. Retuning of the model

can generate a higher-sensitivity version (6.30K) and a lower-sensitivity version (3.95K). Difference between the two versions comes from (1) a choice of empirical function for cloud water phase diagnosis and (2) a treatment of melted cloud ice. In the former, a function similar to Mitchell et al. (1989) results in lower sensitivity while a function similar to Del Genio et al. (1996) results in higher sensitivity. In mid-to-high latitudes, there is a northward shift of cloud water distribution by warming with increase of cloud water in high latitudes. This increase in cloud water is shifted equatorward in lower sensitivity case, compared to that in higher sensitivity case, due to differences in temperature range of empirical function. For the treatment of melted cloud ice, melted cloud ice is diagnosed as cloud water (higher-sensitivity) or as rain (lower-sensitivity). In the latter case, there is less lower-level cloud water in the model climate, which may be the reason for low sensitivity.

The model is used for the 20th century simulation with ‘complete’ forcing both with the lower-sensitivity and higher-sensitivity versions. It is revealed that time-series of annual global mean surface air temperature with lower-sensitivity model fits better to the observed data than that with higher-sensitivity model. Counter-intuitively, the higher-sensitivity model underestimates, rather than overestimates, the historical warming in the 20th century. The higher-sensitivity version seems to be too sensitive to the volcanic forcing. With the simulation

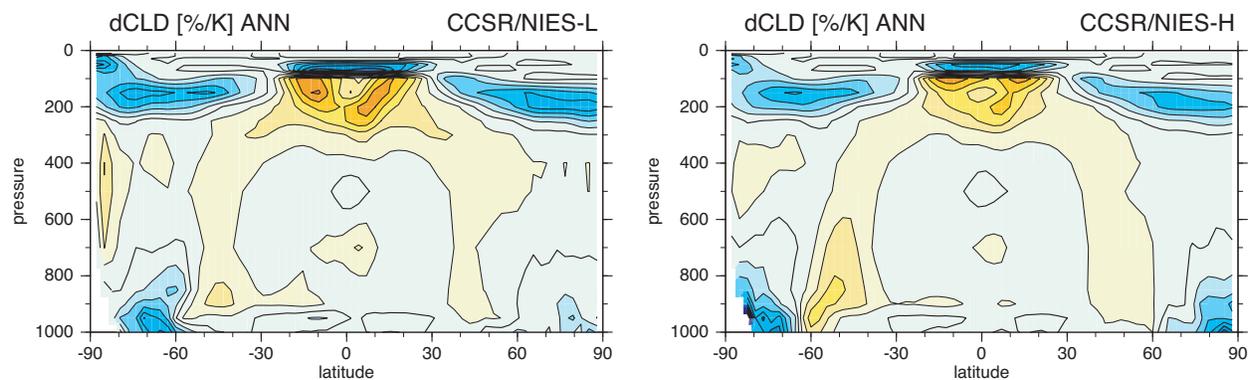


Figure 2. Zonally averaged annual mean cloud amount change normalized by global-mean surface air temperature change for two versions of CCSR/NIES/FRSGC model. (left) lower-sensitivity, (right) higher-sensitivity. Contour interval is 0.5. Blue color denotes increase, while yellow color denotes decrease.

of the Last Glacial Maximum, on the other hand, both versions are within the uncertainty of the sensitivity of low latitude (30°S–30°N) obtained by proxy data.

References

- Del Genio, A.D., M.-S. Yao, W. Kovari and K.K.-W. Lo, 1996: A prognostic cloud water parameterization for global climate models. *J. Climate*, 9, 270-304.
- Emori, S., T. Nozawa, A. Abe-Ouchi, A. Numaguti, M. Kimoto and T. Nakajima, 1999: Coupled ocean-atmosphere model experiments of future climate change with an explicit representation of sulfate aerosol scattering. *J. Meteor. Soc. Japan*, 77, 1299-1307.
- Mitchell, J.F.B., C.A. Senior and W.J. Ingram, 1989: CO₂ and climate: A missing feedback? *Nature*, 341, 132-134.
- Nozawa, T., S. Emori, A. Numaguti, Y. Tsushima, T. Takemura, T. Nakajima, A. Abe-Ouchi and M. Kimoto, 2003: Transient climate change simulations in the 21st century with the CCSR/NIES CGCM under a new set of IPCC scenarios. CGER's Supercomputer Monograph Report, Vol.8, 44pp.
- Takemura, T., H. Okamoto, Y. Maruyama, A. Numaguti, A. Higurashi and T. Nakajima, 2000: Global simulation of aerosol optical thickness distribution of various origins with a three-dimensional model. *J. Geophys. Res.*, 105, 17853-17873.
- Tanaka, T.Y., K. Orito, T. Sekiyama, K. Shibata, M. Chiba and H. Tanaka, 2003: MASINGER, a global tropospheric aerosol chemical transport model coupled with MRI/JMA98 GCM: Model description. *Pap. Meteor. Geophys.*, 53, 119-138.
- Uchiyama, T., A. Noda, S. Yukimoto and M. Chiba, 2004: Study of the estimate of new climate change scenarios based on new emission scenarios. CGER's Supercomputer Activity Report, Vol.11, 41-46.
- Yukimoto, S. and A. Noda, 2002: Improvements of the Meteorological Research Institute global ocean-atmosphere coupled GCM (MRI-CGCM2) and its climate sensitivity. CGER's Supercomputer Activity Report, Vol.10, 37-44.
- Yukimoto, S., A. Noda, A. Kitoh, M. Sugi, Y. Kitamura, M. Hosaka, K. Shibata, S. Maeda and T. Uchiyama, 2001: The new Meteorological Research Institute coupled GCM (MRI-CGCM2) –Model climate and variability. *Pap. Meteor. Geophys.*, 51, 47-88.

Annex 2: Extended Abstracts

Twentieth Century Climate and Probabilistic Estimates of Climate Sensitivity

Reto Knutti

Climate and Environmental Physics, University of Bern, Switzerland

It has recently been suggested that the largely uncertain climate sensitivity can be constrained by relating the reconstructed radiative forcing over the industrial period to the observed surface air warming and the observed ocean heat uptake. The requirement that the modelled warming matches the observed warming should thus place a strong constraint on anthropogenically forced climate models, and the ocean heat uptake should impose an even tighter restriction than the atmospheric warming because of the ocean's large heat capacity [Barnett *et al.*, 2001]. However, these conclusions are hampered by the incompleteness of climate models, by the uncertainty in the surface and ocean temperature records and the uncertainty of the reconstructed radiative forcing over the last 250 years. In particular, it is a problem that reliable data of ocean temperature with sufficient resolution are only available for the last few decades [Levitus *et al.*, 2000]. The variations in the ocean heat content on decadal time scales are large compared to the trend, and current climate models are unable to reproduce the large variability in the observed ocean heat uptake. Further, many radiative forcing components, in particular the aerosol forcing, are still very uncertain.

Figure 1 shows that when climate sensitivity is varied in a simplified climate model [Knutti *et al.*, 2002], the modelled ocean heat uptake and surface warming are consistent with the observed global ocean heat uptake for the period 1955-1995 and the global mean surface warming over the last century for a wide range of climate sensitivities. Even if the radiative forcing is assumed to be known perfectly (Fig. 1a/b), large values for climate sensitivity of more than 6°C cannot be ruled out, since the modelled warming still falls within the observational uncertainty range. The uncertainty in the model ocean mixing parameterisation further widens the possible range of climate sensitivities. If the uncertainties in the radiative forcing are taken into account (Fig. 1c/d), both very low (around 1°C) and very high values cannot be excluded. The main reason is that there is a non-zero probability for the total radiative forcing over the industrial period being zero or even negative [Boucher and Haywood, 2001]. If the total forcing tends to zero, climate sensitivity must be very large to still produce a warming in a climate

model. A significant reduction in the uncertainties of the observational datasets as well as better constraints on the reconstructed radiative forcing are required to considerably reduce the uncertainty in climate sensitivity using this method.

Probabilistic estimates of climate sensitivity can be obtained by large ensembles of simulations with simplified climate models. In this case, climate sensitivity, ocean model parameters and the radiative forcing are varied within their uncertainty ranges, and observations over the last century are used to constrain the ensemble. All recent studies have found that if uncertainties in the observed warming, radiative forcing and the climate models are taken into account properly, the twentieth century warming is insufficient to put a reasonable upper limit on climate sensitivities or to narrow down the probable range assumed so far by IPCC. The derived probability density functions are largely consistent with the IPCC range of 1.5 to 4.5°C [IPCC, 2001], but as long as no expert information is included, a significant probability for climate sensitivity exceeding 4.5°C remains [Andronova and Schlesinger, 2001; Forest *et al.*, 2002; Gregory *et al.*, 2002; Knutti *et al.*, 2002; Knutti *et al.*, 2003].

References

- Andronova, N., and M.E. Schlesinger, Objective estimation of the probability distribution for climate sensitivity, *J. Geophys. Res.*, 106, 22,605-22-612, 2001.
- Barnett, T.R., D.W. Pierce, and R. Schnur, Detection of anthropogenic climate changes in the World's oceans, *Science*, 292, 270-274, 2001.
- Boucher, O., and J. Haywood, On summing the components of radiative forcing of climate change, *Clim. Dyn.*, 18, 297-302, 2001.
- Forest, C.E., P.H. Stone, A.P. Sokolov, M.R. Allen, and M.D. Webster, Quantifying uncertainties in climate system properties with the use of recent climate observations, *Science*, 295, 113-117, 2002.
- Gregory, J.M., R.J. Stouffer, S.C.B. Raper, P.A. Stott, and N.A. Rayner, An observationally based estimate of the climate sensitivity, *J. Clim.*, 15, 3117-3121, 2002.

IPCC, Climate Change 2001: The scientific basis.

Contribution of working group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change, edited by J.T. Houghton, Y. Ding, D.J. Griggs, M. Noguer, B.J. van der Linden, X. Dai, K. Maskell, and C.A. Johnson, pp. 881, Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 2001.

Jones, P.D., M. New, D.E. Parker, S. Martin, and I.G. Rigor, Surface air temperature and its changes over the past 150 years, *Rev. Geophys.*, 37, 173-199, 1999.

Knutti, R., T.F. Stocker, F. Joos, and G.-K. Plattner, Constraints on radiative forcing and future climate change from observations and climate model ensembles, *Nature*, 416, 719-723, 2002.

Knutti, R., T.F. Stocker, F. Joos, and G.-K. Plattner, Probabilistic climate change projections using neural networks, *Clim. Dyn.*, 257-272, DOI 10.1007/s00382-003-0345-1, 2003.

Levitus, S., J.I. Antonov, T.P. Boyer, and C. Stephens, Warming of the world ocean, *Science*, 287, 2225-2229, 2000.

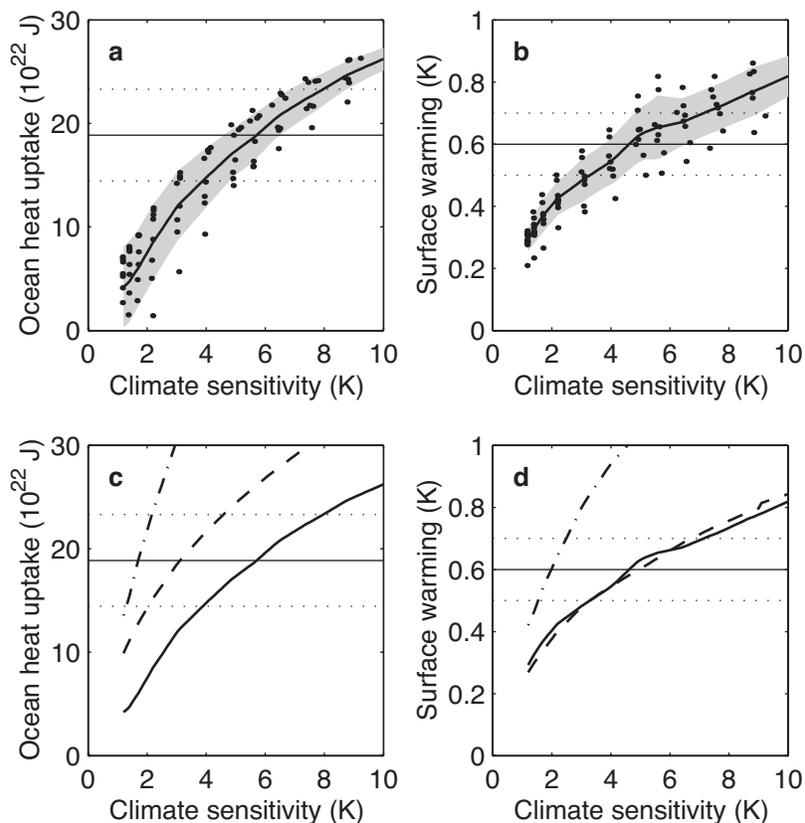


Figure 1. (taken from Knutti *et al.* [2002]) Relation between radiative forcing, climate sensitivity and modelled atmospheric/oceanic warming. (a/b) Global ocean heat uptake 1955 - 1995 (to a depth of 3000 m) and global mean surface air temperature increase 1900 - 2000 versus climate sensitivity (expressed as global mean equilibrium surface temperature increase for a doubling of preindustrial atmospheric CO₂) for eight model setups (different subgrid-scale mixing parameterisations and different vertical diffusivities), using standard reconstructed anthropogenic and natural radiative forcing. Each dot indicates one model simulation. The bold solid curve and shaded band denote the mean and uncertainty (one standard deviation) arising from different ocean mixing properties. Horizontal solid and dotted lines mark the mean and uncertainty (one standard deviation) of the observed ocean heat uptake [Levitus *et al.*, 2000] and observed surface temperature increase [Jones *et al.*, 1999]. (c/d) Model mean values as in (a/b), but when neglecting natural, i. e. solar and volcanic forcings (dashed) or when neglecting the indirect aerosol forcing (dash-dotted). Constraining the climate sensitivity from the observed warming is mainly hampered by uncertainties in the radiative forcing components and temperature data rather than by the range covered by various setups of the climate model used.

Annex 2: Extended Abstracts

Radiative Damping of Annual Temperature Variation: A Proposal for Testing Feedback

S. Manabe¹ and Y. Tsushima²

¹ Princeton University, Princeton, New Jersey

² Frontier Research System for Global Change, Yokohama, Japan

The equilibrium response of global mean surface temperature to radiative forcing is essentially determined by the so-called feedback parameter, which is the rate of radiative damping of the unit anomaly of the global mean surface temperature due to the outgoing radiation from the top of the atmosphere (TOA). Dividing the radiative forcing of climate by the feedback parameter, one gets the radiatively forced, equilibrium response of global surface temperature. The stronger is the rate of radiative damping of surface temperature, the smaller is its equilibrium response to a given radiative forcing.

In the absence of feedback effect, the outgoing radiation at the top of the atmosphere may be approximately equal to the fourth-power of the effective planetary emission temperature, following the Stefan-Boltzmann's law of blackbody radiation. In the actual atmosphere, however, it deviates significantly from the blackbody radiation. For example, when the temperature of the atmosphere increases, its absolute humidity is likely to increase, and enhances the infrared opacity of the atmosphere. Thus, the radiative damping of surface temperature anomaly is weakened, resulting in the increase in the equilibrium response of surface temperature. This is why the water vapor feedback is positive.

The changes in the temperatures of the atmosphere and the Earth's surface affect not only the outgoing long wave radiation but also the reflected solar radiation at the top of the atmosphere. For example, an increase in surface temperature is likely to reduce the area covered by snow and sea ice, thereby reducing the heat loss due to the reflection of incoming solar radiation. Thus, the effective radiative damping of surface temperature anomaly is reduced, enhancing the sensitivity of climate.

According to the third IPCC (2001) report, the previously estimated range of the equilibrium response of the global mean surface temperature to the doubling of atmospheric carbon dioxide has not reduced substantially over the last decade and remains between 1.5 and 4.5°C. Clearly, the large range in the estimated sensitivity of surface temperature is attributable in no small part to our inability to reliably determine the influence of feedback upon the radiative damping of surface temperature anomaly.

In the present study, we estimate the rate of the radiative damping of the annual variation in the global mean surface temperature. Using the TOA fluxes of radiation obtained from the Earth radiation Budget Experiment (ERBE), we evaluate how the overall feedback of the atmosphere alters the radiative damping of the annually varying anomaly of surface temperature. Specifically, we compute the gain factor, i.e., a factor that represents the contribution of the overall feedback to the reduction of the radiative damping of the global surface temperature anomaly. To identify the systematic bias of the overall feedback simulated by a model, the gain factor thus estimated is then compared with the gain factor of the feedback simulated by the model.

The annual variation of the global mean surface temperature is attributable mainly to the difference in effective thermal inertia between continent and ocean rather than to the annual variation of global mean, incoming solar radiation that is very small. Because the seasonal variation of surface temperature is much larger over continents than over oceans, the annual variation of the global mean surface temperature is dominated by the contribution from the continents in Northern Hemisphere. Its annual range is about 3.3°C with highest temperature in July and the lowest in January. The range is comparable in magnitude to a current estimate of the equilibrium response of global mean surface temperature to the doubling of CO₂ concentration in the atmosphere.

Since the pattern of the annual variation of surface temperature (see, for example, Figure 1b of Tsushima and Manabe, 2001) differs greatly from that of global warming simulated by a model, it is quite likely that the rate of the radiative damping of the global mean surface temperature anomaly is significantly different between the two. As noted by Raval and Ramanathan (1989) and Inamdar and Ramanathan (1998), the rate of radiative damping of local surface temperature is similar to that of global mean temperature under clear sky. Therefore, it is likely that the rate of the radiative damping of global surface temperature anomaly is similar between the annual variation and global warming. On the other hand, a similar statement may not be made for the albedo- and water vapor-feedback. Nevertheless, we decided to use the annual variation for a rough estimate of the gain

factor of the overall feedback because it is the largest climatic change one can observe. The availability of data from the ERBE is another decisive factor for conducting the analysis presented here.

In the present study, we estimated the rate of radiative damping of annually varying anomaly of the global mean surface temperature, using the TOA-fluxes of terrestrial radiation and reflected solar radiation obtained from ERBE. We found that the radiative damping is reduced by as much as 70% because of the overall feedback, and is only 30% of what is expected for the black body with the planetary emission temperature. In other words, the gain factor of the overall feedback turned out to be as large as 0.7.

Given that the radiative forcing of the CO₂-doubling is 4 Wm⁻² (e.g., Hansen et al., 1997) the equilibrium response of the global mean surface temperature to the doubling would be about 1.2°C in the absence of feedback. If one assume that the gain factor of the annual variation obtained here (i.e., $f = 0.7$) were applicable to global warming, the equilibrium response to CO₂-doubling in the presence of feedback would be 4.0°C, which is about 3.3 ($= 1 / (1-f)$) times as large as 1.2°C, (i.e., the equilibrium response in the absence of feedback). Because of the difference between the pattern of the annual variation and that of global warming, the gain factor of the albedo feedback is likely to be larger for the annual variation than for global warming. Therefore, the equilibrium response of the global mean surface temperature to the doubling of atmospheric carbon dioxide may be less than 4.0°C ($= 1.2^\circ\text{C} / [1-0.7]$), and is near the middle of the climate sensitivity range as estimated by IPCC (i.e., 1.5°C~4.5°C).

Similar feedback analysis is applied to the annual variation obtained from the three general circulation models (submitted to AMIP I), in which the microphysical properties of cloud is computed explicitly. Although the gain factors of the overall feedback in these models happens to be approximately similar to the gain factor (Figure 1), which is determined using ERBE, the long wave and solar gain factors obtained from these models are quite different from the observed (Fig. 1). Since the difference almost disappears if the contribution from the cloud feedback is removed, we believe that a major fraction of the discrepancy is attributable to the failure of the models to satisfactorily simulate the long wave and solar components of the cloud feedback (Tsushima and Manabe, 2001). To evaluate satisfactorily the feedback of a model, it is therefore desirable to compute not only the gain factor of the overall feedback but also those of relevant feedbacks that operate on the annual variation, and compare them with observation.

Proposed analysis of simulated feedbacks

As noted above, it is probable that the gain factors of the overall feedback are significantly different between the annual variation and global warming. Nevertheless, it is likely that they are positively correlated to each other, because of the similarity in the the physical mechanisms involved. One can therefore compute the gain factors of the overall feedback for the annual variation, compare it with the simulated gain factor, and estimate the bias of the overall feedback in climate models used for the projection of global warming as we did in the present study.

For example, one can conduct regression analysis between the gain factor of global warming and that of

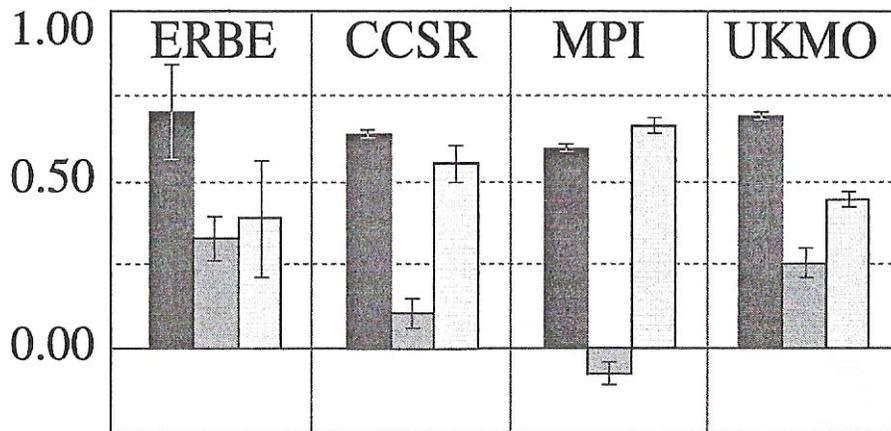


Figure 1. Gain factors from the ERBE observation and old versions of the three models, which were constructed at Center for Climate System Research/National Institute for Environment Study (CCSR/NIES), Max-Planck Institute for Meteorology (MPI), and United Kingdom Meteorological Office (UKMO) and were submitted to AMIP-I model intercomparison project (Gates, 1992). Black bar represents the gain factor of the overall feedback. Dark and light grey bars represent solar and terrestrial gain factors, respectively. Note that standard errors are indicated by the line segments attached to the bars identified above. The line segments attached to these bars indicate standard error. Here, gain factor is defined as the factor that represents the reduction in the radiative damping due to the overall feedback. (Hansen et al., 1984; see also, Wetherald and Manabe, 1988).

the annual variation obtained from many climate models submitted to Intergovernmental Panel on Climate Change. On the regression line through the scatter plots, each of which represents a set of the two gain factors, one can seek the most likely value of gain factor for global warming, referring to the gain factor of the annual variation obtained from ERBE.

It is very desirable to conduct similar regression analysis for individual components of feedback (i.e., water vapor feedback, albedo feedback, and the cloud feedback) using the TOA fluxes of clear- and whole skies obtained from ERBE.

References

- Gates, W. L., 1992: AMIP: The Atmospheric Model Intercomparison Project. *Bulletin of American Meteorological Society*, 73, 1962-1970.
- Hansen, J. et al., 1984: Climate sensitivity: Analysis of feedback mechanisms. In *Climate Process and Climate Sensitivity*, edited by J. Hansen and T. Takahashi, *Geophysical Monograph Series*, vol. 29, 130-163, American Geophysical Union, Washington, D. C.
- Hansen, J. M. Sato, R, Ruedy, 1997: Radiative forcing and climate response. *J. Geophys Res*, **102(D6)**, 6831-6864.
- Inamdar, A.K., and V. Ramanathan, 1998: Tropical and global scale interaction among water vapor, atmospheric greenhouse effect, and surface temperature. *J. Geophys. Res.*, **103**, 32177-32194.
- Raval, A., and V. Ramanathan, 1989: Observational determination of greenhouse effect. *Nature* **342**: 758-761.
- Tsushima, Y., and S. Manabe, 2001, Influence of cloud feedback on annual variation of global mean surface temperature, *Journal of Geophysical Research (Atmosphere)* **106(D19)**, 22, 635-22,646, (October 16, 2001).
- Tsushima, Y., S. Manabe, and A. Abe-Ouchi, 2003: Radiative damping of annual temperature variation: Observational constraint on simulated feedback. submitted to *Climate Dynamics*.
- Wetherald, R. T., and S. Manabe, 1988: Cloud feedback process in a general circulation model, *Journal of Atmospheric Sciences*, **45**, 1397-1415.

Annex 2: Extended Abstracts

Ocean Heat Uptake and Climate Sensitivity in Five Coupled Models

Gerald A. Meehl

National Center for Atmospheric Research

Five global coupled climate models are compared for their climate sensitivity characteristics related to ocean heat uptake. To a greater degree than the other components, the atmospheric model “manages” the relevant global feedbacks including ice/albedo, water vapor and clouds (Meehl et al., 2004). The atmospheric model also affects the meridional overturning circulation in the ocean, as well as the ocean heat uptake characteristics. This is due to changes in surface fluxes of heat and fresh water that affect surface density in the ocean. For global sensitivity measures, the ocean, sea ice and land surface play secondary roles, even though differences in these components can be important for regional climate changes. Two models with the same atmosphere and sea ice components but different ocean (PCM and PCTM) have the most similar response to increasing CO₂, followed closely by CSM with the same atmosphere and different ocean and sea ice from either PCM or PCTM. The CCSM has a different response from either of the other three, and in particular is different from PCTM in spite of the same ocean and sea ice but different atmospheric model components. CCSM3 differs mainly from CCSM in regards to physics in the atmosphere, and consequently has a different response yet again, consistent with the idea that changes in the atmosphere have the greatest effect on overall climate sensitivity, including ocean response (Fig. 1).

Ocean heat uptake efficiency is shown to depend on the specific characteristics of the atmosphere in particular, with a change in the atmospheric model (from PCTM to CCSM) reducing not only the strength of the MOC in the control run, but also affecting the heat uptake efficiency with increased CO₂ (Fig. 2). The change from CCSM to CCSM3 has the opposite effect. That is, the CCSM, with the lowest climate sensitivity, has a somewhat greater percent reduction of MOC compared to the PCTM. However, the CSM has a slight strengthening of MOC, greater high latitude warming, but less heat uptake efficiency. The CCSM3 has greater sensitivity than CCSM, with almost twice the percent change (weakening) of the THC (Fig. 3).

Therefore, the nature of the climate system response to an input of increased energy to the system from increased CO₂ can be partitioned in various ways between atmosphere and ocean, with the nature of that partitioning related to the specific characteristics of the feedbacks in the system mainly having to do with the atmospheric model.

Reference

Meehl, G.A., W.M. Washington, J.M. Arblaster, and A. Hu, 2004: Factors affecting climate sensitivity in global coupled models. *J. Climate*, 17, 1584–1596.

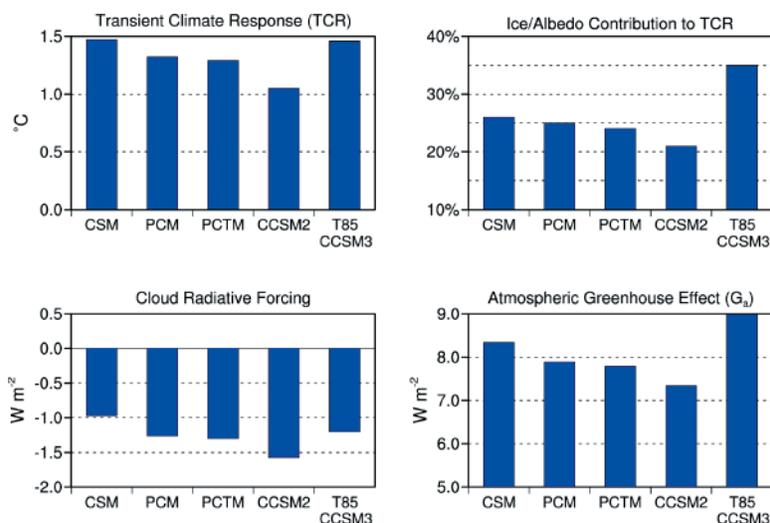


Figure 1. a) Globally averaged surface air temperature difference (C) around the time of CO₂ doubling for years 61-80 minus control (transient climate response or TCR); b) ice-albedo feedback contribution (%) to TCR from the five models; c) as in (a) except for cloud radiative forcing (Wm⁻²); d) as in (a) except for atmospheric greenhouse effect (Wm⁻²).

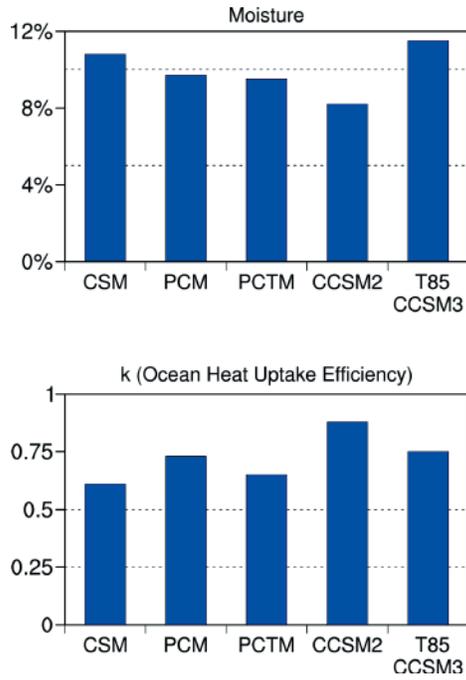


Figure 2. a) Same as Fig. 1a except for percent change in total column moisture, b) same as (a) except for ocean heat uptake efficiency.

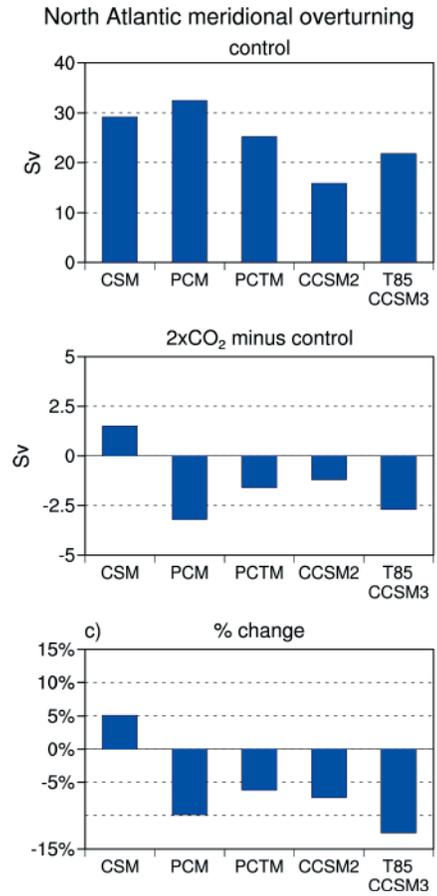


Figure 3. North Atlantic Ocean meridional overturning circulation (MOC) a) maximum values from the control run for the five models (Sv), b) differences of MOC around the time of CO₂ doubling, years 61-80, minus respective control run averages for the five models (Sv), and c) percent change in MOC for a doubling of CO₂ (values in (b) divided by values in (a)).

Probabilistic Estimates of Climate Sensitivity from GCM Ensembles

James Murphy, David Sexton, Matthew Collins, Glen Harris and Mark Webb

Hadley Centre, Met Office, FitzRoy Road, Exeter EX1 3PB, UK

For a given scenario of external forcing GCM projections of climate change are subject to uncertainties arising from modelling errors and the effects of natural variability. In principle these uncertainties can be estimated by constructing large ensembles of simulations which sample alternative representations of Earth system processes and alternative initial states. In practice resource limitations have prevented this so we typically rely on pooling individual simulations from different GCMs to create small ensembles from which uncertainties are estimated (e.g. Cubasch et al 2001, Palmer and Räisänen 2002). These provide useful information but do not provide a credible basis for the construction of probabilistic projections because (a) the ensembles are too small, (b) they are not constructed to sample uncertainties in a systematic manner (Allen and Ingram 2002) and (c) they have not been accompanied by objective measures of reliability which can be used to weight the projections of different GCMs.

We report a first step towards addressing these issues (See Murphy et al (2004) for a fuller description). Simulations of the equilibrium response to doubled CO₂ were made using 53 versions of the Hadley Centre GCM coupled to a mixed layer ocean. Each ensemble member differed from the standard version (HadAM3, see Pope et al, 2000) by a perturbation to one of 29 poorly-constrained parameters influencing various surface and atmospheric physical processes. A number of parameters were found to influence climate sensitivity, however it was not possible to construct a probability distribution directly from the ensemble results because the effects of multiple parameter perturbations, representing interactions between uncertainties in different processes, were not sampled. However, the impact of such interactions can be estimated by assuming that the effects of individual parameter perturbations combine linearly. This allows us to predict the results of a much larger (multi-million member) ensemble of GCM versions containing multiple parameter perturbations sampled by assuming a uniform prior for each of the 29 parameters within limits estimated by experts.

If we assume that all model versions are equally plausible we obtain a probability density function (pdf) for climate sensitivity with a 5-95% probability range of 1.9-5.3 Deg C. However we have also developed a "Climate Prediction Index" (CPI) for the purpose of weighting model versions according to reliability.

The CPI is currently determined from normalised error variances of simulated climatological patterns of 32 surface and atmospheric variables. Weighting the simulations of GCM versions according to the CPI changes the 5-95% probability range to 2.4-5.4 Deg C.

Previously pdfs of sensitivity have been estimated by exploring the range of predictions of simpler climate models consistent with observed changes in the forcing and response of climate since the industrial revolution (Andronova and Schlesinger 2001, Forest et al 2002, Gregory et al 2002, Knutti et al 2002). Our approach differs from the above studies and allows us to estimate a pdf for sensitivity determined by both the complex physical interactions built into the GCM and the fit of different model versions to a wide range of present day observations. However it is subject to a number of caveats, including the following:

- (1) We consider only uncertainties in surface and atmospheric physical processes, neglecting uncertainties associated with feedbacks involving the ocean circulation, the terrestrial biosphere, atmospheric chemistry and ocean biogeochemistry.
- (2) Perturbing uncertain parameters in a single GCM does not, in any case, sample the full range of possible modelling uncertainties. In particular, "structural" differences between GCMs developed at different centres (different choices of resolution, variations in the basic nature of parameterisation schemes, variations in the range of parameterisations included) are not sampled.
- (3) The results depend on the assumed *a priori* distributions for parameter values, in particular on their expert-specified extremes.
- (4) Errors arise from the assumption that the effects of individual parameter perturbations combine linearly.
- (5) The estimated reliability of alternative model versions depends on the chosen metric of model skill, the set of verifying observations on which it is based, the relative weights attached to different observations and the assumptions required to convert values of the metric into estimates of relative likelihood.

These caveats reflect the fact that ensemble climate prediction is in its infancy. Addressing them will involve a number of substantial developments. For example (1) would require ensemble simulations carried out with

full Earth System Models, while (2) would require coordinated ensemble experiments involving systematic exploration of process uncertainties in a number of different GCMs; (3) could be assessed by comparing parallel ensembles generated using alternative priors for uncertain parameters and (4) requires the production of very large ensembles which sample parameter space more comprehensively (Stainforth et al 2004). Regarding (5), further research is needed to determine the best methods of relating uncertainties in verifiable aspects of GCM simulations to uncertainties in their projections of future changes (see accompanying abstract by Senior et al).

Ultimately the assessment of climate-related risks requires provision of probabilistic projections for time-dependent changes. For example carbon cycle feedbacks may be relatively unimportant as a source of uncertainty in equilibrium climate sensitivity, yet may be a key source of uncertainty in the transient response to realistic emissions scenarios (e.g. Cox et al 2000). Furthermore, impact assessments typically require uncertainties specified at a regional level, rather than for idealised global quantities such as climate sensitivity. Our results suggest that only a comprehensive approach to sampling modelling uncertainties can provide a realistic basis for quantifying the range of possible regional changes (Murphy et al 2004, Stainforth et al 2004). It is therefore essential that the design of GCM ensembles samples uncertainties relevant to regional changes in impact-related variables such as precipitation, storm intensity and soil moisture content and that climate metrics are designed which are capable of constraining projections of regional changes.

References

Allen, M. R. & Ingram, W. J. (2002). Constraints on future changes in climate and the hydrological cycle. *Nature* **419**,224-232.

- Andronova, N. G. & Schlesinger, M. E. (2001). Objective estimation of the probability density function for climate sensitivity. *J. Geophys. Res.* **106**,22605-22612.
- Cubasch, U. et al. in *Climate Change 2001, The Science of Climate Change* Ch. 9 (eds Houghton, J. T. et al.) 527-582 (Cambridge U. Press, Cambridge, 2001).
- Forest, C. E., Stone, P.H., Sokolov, A. P., Allen, M. R. & Webster, M. D. (2002). Quantifying uncertainties in climate system properties with the use of recent climate observations. *Science* **295**,113-117.
- Knutti, R., Stocker, T. F., Joos, F. & Plattner, G. K. (2002) Constraints on radiative forcing and future climate change from observations and climate model ensembles. *Nature* **416**,719-723.
- Gregory, J. M., Stouffer, R. J., Raper, S. C. B., Stott, P. A. & Rayner, N. A. (2002) An observationally based estimate of the climate sensitivity. *J. Clim.* **15**,3117-3121.
- Murphy J M. et al. (2004) Quantification of modelling uncertainties in a large ensemble of climate change simulations. *Nature*, to appear.
- Palmer, T. N. & Räisänen, J. (2002). Quantifying the risk of extreme seasonal precipitation events in a changing climate. *Nature* **415**,514-517.
- Pope, V. D., Gallani, M., Rowntree, P. R. & Stratton, R. A. (2000). The impact of new physical parameterisations in the Hadley Centre climate model – HadAM3. *Clim. Dynam.* **16**,123-146.
- Senior C. A. (2004) Report on the joint WCRP CFMIP/IPCC expert meeting on “Climate Sensitivity and Feedbacks: Summary and Recommendations”, Exeter, UK, April 19-22.
- Stainforth, D. A. et al. (2004). Evaluating uncertainty in the climate response to changing levels of greenhouse gases. Submitted to *Nature*.

Annex 2: Extended Abstracts

Climate Sensitivity of the Last Glacial Maximum from Paleoclimate Simulations and Observations

Bette Otto-Bliesner, Esther Brady, and Zav Kothavala

National Center for Atmospheric Research, Boulder, Colorado, US

Global coupled climate models run for future scenarios of increasing atmospheric CO₂ give a range of response of the global average surface temperature. Regional responses, including the North Atlantic overturning circulation and tropical Pacific ENSO, also vary significantly among models. The second phase of the Paleoclimate Modeling Intercomparison Project (PMIP 2) is coordinating simulations and data syntheses for the Last Glacial Maximum (21,000 years before present [21 ka]) to allow an independent assessment of climate sensitivity.

Atmospheric CO₂ concentrations at the Last Glacial Maximum (LGM) have been estimated using measurements from ice cores to be 185 ppmv, approximately 50% of present-day values. Global, annual mean surface temperature simulated by the T42 slab ocean version of CCSM3 shows a cooling of -2.8°C for LGM CO₂ levels and a warming of 2.5°C for a doubling of CO₂. The climate sensitivity factors calculated from these two simulations are 0.8 K/Wm^{-2} for LGM and 0.7 K/Wm^{-2} for $2\times\text{CO}_2$. At LGM atmospheric CH₄ concentration is estimated to be 350 ppbv, N₂O concentration to be 200 ppbv, and chlorofluorocarbons were not present. Large ice sheets covered North America and Eurasia and sea level was reduced by approximately 105 m. When these additional forcings are included in the T42 slab ocean version

of CCSM3, global surface temperature cools 5.7°C compared to present. A T42 fully coupled CCSM3 simulation for LGM shows comparable, but somewhat larger, cooling of global surface temperature of 5.9°C .

Regional signatures of the climate system to changed LGM forcing are also an important measure of climate sensitivity. Understanding and comparing the response of the Atlantic meridional overturning circulation (MOC) to LGM forcing in coupled atmosphere-ocean climate models and in proxy observations is a component of PMIP-2. CCSM3 shows a weakening and shallowing of the Atlantic meridional overturning circulation at LGM compared to present. A CSM1 simulation for LGM gives a similar tendency. CCSM3, though, with a more realistic MOC for present gives maximum overturning of 13 Sv at LGM compared to 21 Sv in the CSM1 LGM simulation. Previous water-hosing experiments with CSM1 suggest that the Atlantic MOC is sensitive to the initial state. Future simulations with CCSM3 and other PMIP-2 coupled models will explore this sensitivity.

In this talk, the climate sensitivity and the response of the Atlantic MOC and tropical Pacific Ocean for the NCAR climate models, CSM1 and CCSM3, to LGM forcing will be discussed. Results of coupled ocean-atmosphere simulations from other centers and proxy observations will also be included.

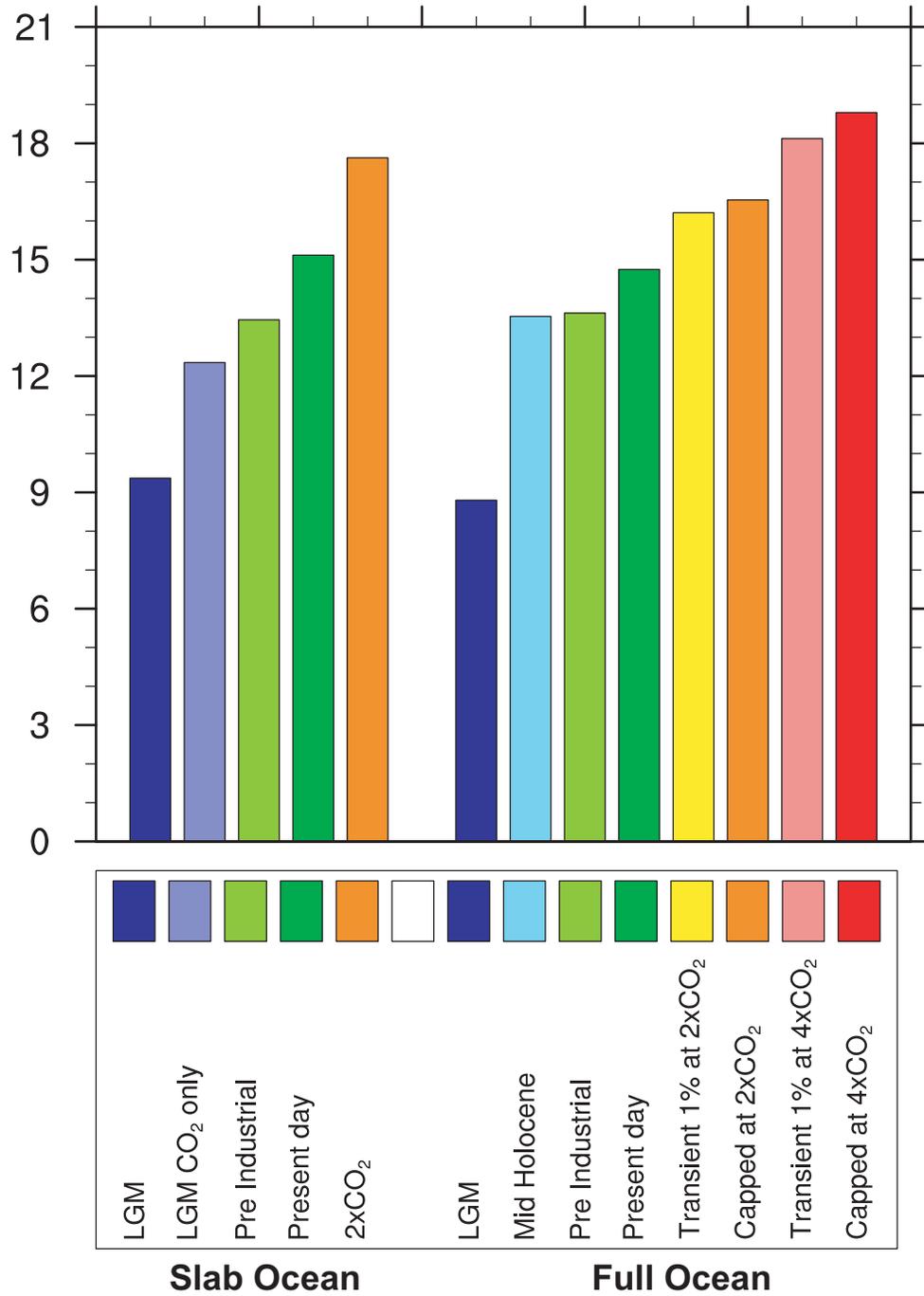


Figure. Global, annual surface temperature simulated by slab (left) and full (ocean) versions of CCSM3 at T42 resolution for past, present, and future climates.

Annex 2: Extended Abstracts

Paleoclimatic Evidence for Future Greenland Ice Sheet Instability and Rapid Sea Level Rise

Jonathan T. Overpeck¹, Bette L. Otto-Bleisner², Jeffrey T. Kiehl², Gifford H. Miller³, and IGBP PAGES CAPE Project Members

¹ *Institute for the Study of Planet Earth and Department of Geosciences, University of Arizona, Tucson, AZ 88721, USA; jto@u.arizona.edu*

² *National Center for Atmospheric Research, P.O. Box 3000, Boulder, Colorado 80307, USA*

³ *INSTAAR and Department of Geological Sciences, University of Colorado, Boulder, CO 80309, USA*

The peak of the last interglacial (LIG) period, starting $129,000 \pm 1000$ years ago, provides the most recent paleoclimatic glimpse into how the high northern latitudes respond to significant warming above present day. Changes in the Earth's orbit (e.g., the astronomical theory of climate change) resulted in significant positive warm-season insolation anomalies at this time. These anomalies were significantly larger than the peak positive anomalies associated with climate change early in the present Holocene interglacial, and for this reason it is not surprising that LIG warmth apparently had a much larger impact on the arctic than Holocene warmth. Most importantly, it appears that LIG warmth was sufficient to melt much of the Greenland Ice Sheet. Well-dated corals above present sea level provide unequivocal evidence that sea level was 3 to 6 meters above present at the peak of the LIG period (Israelson and Wohlfarth 1999; McCulloch and Esat 2000; Muhs 2002; Shackleton, Sanchez-Goni et al. 2003), and the failure to drill structurally intact ice older than this time over much of Greenland supports the contention that GIS melting was the cause of this sea water rise (Cuffey and Marshall 2000).

The goal of our work is to combine the use of paleoenvironmental data with a new modeling study to reevaluate the sensitivity of the arctic, and Greenland Ice Sheet in particular, to possible future arctic warmth comparable to that of the LIG period. We used a state-of-the-art coupled atmosphere ocean climate model to simulate the climate of 130,000 years ago, as well as the climate of the next 140 years. The model is the National Center for Atmospheric Research Community Climate System Model [CCSM], Version 2). CCSM2 includes an atmospheric model at resolution of 3.75 latitude and longitude and 26 vertical levels, an ocean model with 320x384 gridpoints and 40 levels, a sea ice model which includes both thermodynamics and elastic-viscous-plastic dynamics, and a land model that includes explicit river runoff and the physical effects of vegetation and land ice. Vegetation and land ice coverage are prescribed at their present-day distributions and are unchanged

for the LIG simulations. In accordance with what is known about climate forcing at the peak of the LIG, we prescribed our experimental forcing to be the same as 1990 with the exception that seasonal insolation and atmospheric greenhouse levels ($\text{CO}_2 = 280$ ppmv, $\text{CH}_4 = 600$ ppbv) were set to be the same as 130,000 years ago. Although peak warmth of the LIG could have occurred 1-2000 years after 130,000 yr B.P., the forcing would not have been significantly different from that we prescribed. We used the same model to simulate into the future, setting atmospheric CO_2 to increase 1% per year from 1990 values of 355 ppmv.

Results reveal several key aspects of the LIG climate. First, we confirm that sea level rise at ca. 130,000 years ago must have been associated with Greenland Ice Sheet melting rather than melting in the Antarctic. This fact is not surprising given the large positive summertime insolation anomaly at high northern latitudes at that time (nearly 70 W/m^2 in early summer), and the lack of any significant warm (i.e., to above freezing) anomaly in south polar latitudes. Our simulation (Fig. 1) suggests that the summer temperatures needed to melt the GIS were on the order of 3 to 5°C above present (paleoclimate data from the Arctic support the higher end of this range), and that these temperatures were associated with a net reduction in snowfall over Greenland as well. Not surprisingly, simulated summer sea ice was greatly reduced at 130,000 yr B.P.

Comparison of the summer season warmth sufficient to have melted the GIS 130,000 years ago with simulated future climate (Fig. 1) indicates that the high northern latitudes around Greenland will be as warm as 130,000 years ago, and hence warm enough to melt the GIS, sometime before AD 2130, and when the CO_2 content of the atmosphere reaches a level three to four times the pre-industrial level. As with our paleoclimate LIG simulation, it does not appear likely that increased snowfall or ocean circulation changes will retard GIS melting. Our results also suggest that the current Arctic warming, ice sheet melting and accelerating retreat of

summer Arctic sea ice thickness and extent could be a harbinger of sea ice retreat similar to that associated with GIS melting around 130,000 years ago.

A closer look at past ice sheet melting *rates* yields even more startling insight into potential future sea level rise. Sea level due to ice sheet (excepting Greenland) melting averaged 11mm/year over a large part of the last deglaciation between 13,800 to 7000 years ago (Bard, Hamelin et al. 1996). The penultimate deglaciation culminating with the LIG sea level highstand 3-6m above present day, however, was driven by a significantly larger high latitude summertime insolation anomaly, and was also able to melt much of Greenland. The most realistic sea level rise scenario for this penultimate deglaciation, also based on careful dating of sea-level dependent corals, suggests that GIS melting and associated sea level rise occurred much more rapidly, with estimated rates as high as 20 mm/yr or even higher (30 to 50 mm/yr; (McCulloch and Esat 2000)). This rate of sea level rise makes sense relative to the last deglaciation and its smaller summer insolation maximum, and indicates that much of the GIS could melt in just a few centuries. It is not surprising that glaciologists have recently begun to unravel new processes by which ice sheets can collapse faster than previously thought. High levels of atmospheric pollution (i.e., soot) may also hasten ice sheet melting even more than in the past (Hansen and Nazarenko 2004), and any sea level rise related to the collapse of the GIS will be on top of sea level rise due to ocean warming and melting of other land ice.

References

- Bard, E., B. Hamelin, et al. (1996). “Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge.” *Nature* **382**(6588): 241-244.
- Cuffey, K. and S. Marshall (2000). “Substantial contribution to sea-level rise during the last interglacial from the Greenland ice sheet.” *Nature* **404**: 591-552.
- Hansen, J. and L. Nazarenko (2004). “Soot climate forcing via snow and ice albedos.” *Proceedings of the National Academy of Sciences of the United States of America* **101**(2): 423-428.
- Israelson, C. and B. Wohlfarth (1999). “Timing of the last-interglacial high sea level on the Seychelles Islands, Indian ocean.” *Quaternary Research* **51**(3): 306-316.
- McCulloch, M. T. and T. Esat (2000). “The coral record of last interglacial sea levels and sea surface temperatures.” *Chemical Geology* **169**(1-2): 107-129.
- Muhs, D. R. (2002). “Evidence for the timing and duration of the last interglacial period from high-precision uranium-series ages of corals on tectonically stable coastlines.” *Quaternary Research* **58**(1): 36-40.
- Shackleton, N. J., M. F. Sanchez-Goni, et al. (2003). “Marine Isotope Substage 5e and the Eemian interglacial.” *Global and Planetary Change* **36**(3): 151-155.

Simulated summer (JJA) surface air temperature anomalies

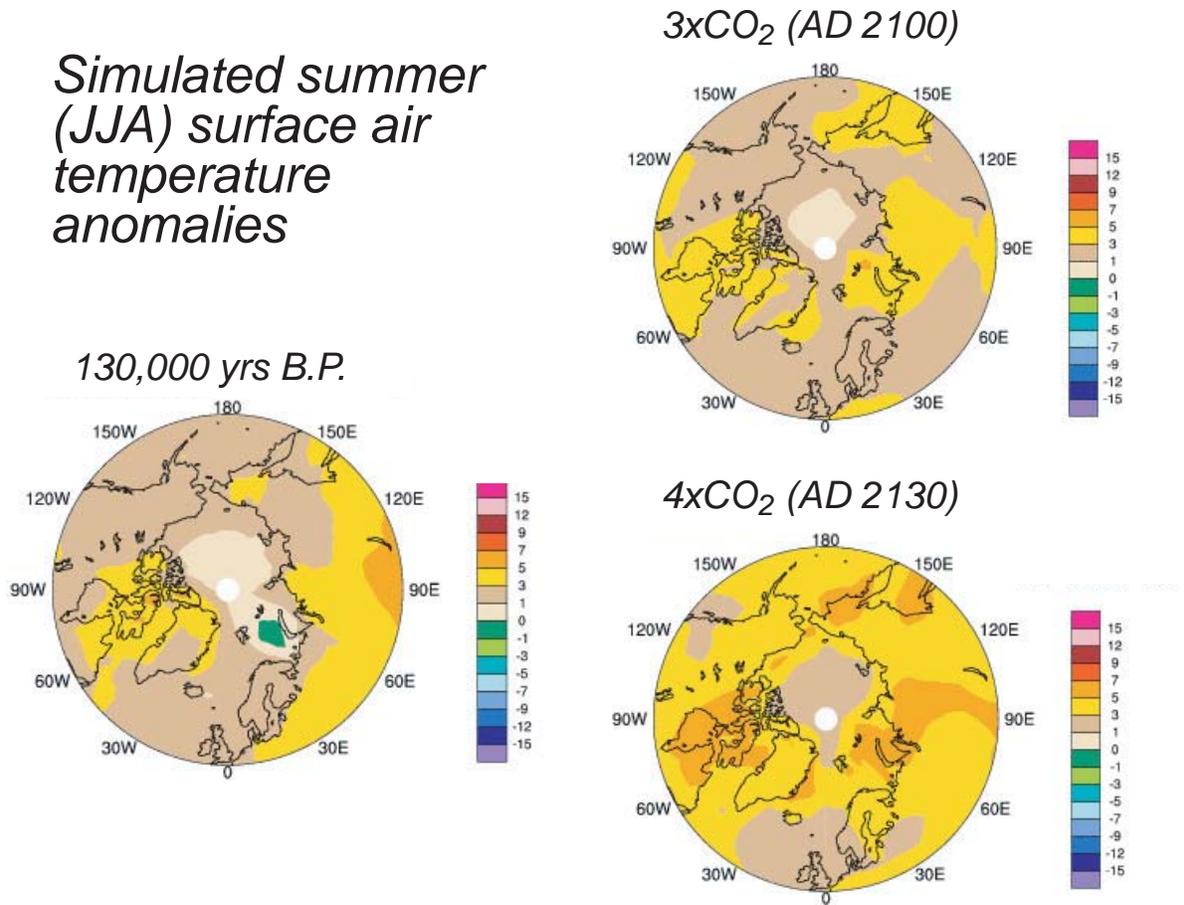


Figure 1. Simulated summer (JJA) surface air temperature anomalies for 130,000 years ago, 2100 AD and 2130 AD (assuming a 1% per year increase in CO₂ starting from a 1990 value of 355 ppmv). Results indicate that future warming comparable to 130,000 years ago be warm enough to melt the Greenland Ice Sheet at a rates comparable to those that characterized the penultimate deglaciation (i.e., >2 meters of sea-level equivalent per century).

Annex 2: Extended Abstracts

Probability Distributions of CO₂-Induced Global Warming as Inferred Directly from Multi-Model Ensemble Simulations

Jouni Räisänen

Department of Physical Sciences, Division of Atmospheric Sciences, P.O.B 64, FIN-00014 University of Helsinki, Finland
(Email: jouni.raisanen@helsinki.fi)

The most straightforward way to estimate the probability distribution of CO₂-induced global warming is to directly use results of different climate models. This approach has an obvious (and big) caveat: climate changes in the real world may not necessarily follow model results (e.g., Allen and Ingram 2002). However, this simple method of estimating probability distributions provides a natural benchmark against which to compare the results of more rigorous methods.

Two strategic numbers associated with the global mean temperature are studied: (i) the transient climate response (TCR) defined as the warming that occurs at the time of doubling of CO₂ when CO₂ doubles in 70 years, and (ii) the equilibrium warming resulting from the doubling of CO₂, widely known as climate sensitivity (CS). The TCR is evaluated for 20 models participating in the CMIP2 intercomparison (Meehl et al. 2000). For CS, the 15 model results given in Table 9.1 of Cubasch et al. (2001) are used. TCR is evaluated for full atmosphere-ocean GCMs, CS for atmospheric GCMs coupled to a mixed-layer ocean.

The 20 values for TCR are in ascending order 1.0, 1.1, 1.3, 3×1.4, 6×1.6, 1.8, 2×1.9, 3×2.0, 2.1 and 3.1°C, and the 15 values for CS 2.0, 2.1, 2.2, 3.1, 3.2, 3.3, 3.4, 3.5, 2×3.6, 3.7, 4.1, 4.3, 4.8 and 5.1°C. Let us assume that these values are a random sample of

some underlying probability distribution, neglecting for the moment the fact that the different models are not necessarily independent from each other. What can we infer of the properties of this hypothetical underlying distribution?

First, one needs to guess the form of the probability distribution. I only consider two candidates: a normal and a log-normal distribution. Visual comparison of the fitted distributions with the original data (Fig. 1) suggests that the TCR data are not well described by a normal distribution. This is confirmed by statistical tests, which show that the original TCR distribution is both positively skewed and leptocurtic at 98% (two-sided) significance level. The log-normal distribution provides a better fit, with no statistically significant deviations from the original data. In the case of CS both the two distributions fit the data within the uncertainty caused by the limited sample size.

Thus, if TCR and CS are to be described by the same form of an analytical distribution, the log-normal distribution appears to be a better candidate than the normal one. However, there is no way to rule out the possibility that the actual underlying distributions are of some other, unknown form. It may also be noted that the significant skewness and curtosis in the TCR distribution both result from one model (CCSR/NIES2; Nozawa et

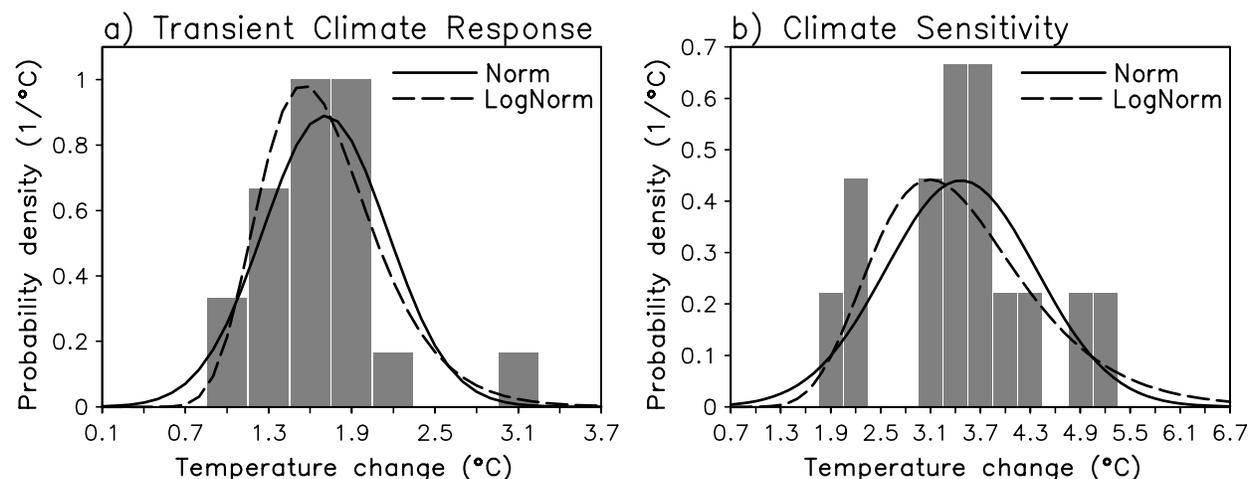


Figure 1. Normal (solid line) and log-normal (dashed line) probability distributions fitted to (a) TCR in 20 models and (b) CS in 15 models. The bars show the original data with a bin width of 0.3°C.

al. 2000) that simulates substantially larger warming than any of the others.

The 5th, 50th and 95th percentiles of the derived probability distributions are given in Table 1. The best-estimate 5-95% uncertainty range for TCR is 1.1-2.5°C, and that for CS 2.1-5.3°C, when using the log-normal distribution. For the normal distribution, both the 5th and 95th percentiles are slightly lower. The median for CS (about 3.4°C) is twice the value for TCR (1.7°C), even though it should be noted that the two distributions have been derived using slightly different sets of models. The widely cited range (1.5-4.5°C) and best estimate (commonly given as 2.5°C) for CS are somewhat on the lower side of the model results.

Because the derived analytical probability distributions are based on a limited number of model results, they suffer from sampling uncertainty. This source of uncertainty was estimated by first generating a large number of artificial 15- and 20-model random samples by using the parameters of the best-fit normal or log-normal distributions, and by then repeating the distribution fitting for each of these samples. At least for the cases considered in Table 1, the sampling uncertainty is larger than the differences between the normal and the log-normal distributions. The best-estimate 5%, 50% and 95% quantiles for the normal distribution are always within the 5-95% sampling range for the log-normal distribution, and vice versa. In the case of the log-normal fit, the sampling uncertainty is largest in the upper end of the distribution; note in particular the large (4.39-6.35°C) uncertainty in the 95th percentile of CS. For the normal distribution fit, both the lower and upper ends suffer from larger sampling uncertainty than the middle of the distribution.

Table 1. Quantiles of TCR and CS, as derived from the fitted analytical probability distributions. In each table entry, the first value gives the best estimate and the next two (in parentheses) the 5-95% sampling uncertainty inferred from Monte Carlo simulations.

	Quantile	Normal Distribution	Log-Normal Distr.
TCR	5%	0.97 (0.72-1.23)	1.09 (0.95-1.27)
	Median	1.71 (1.55-1.87)	1.66 (1.51-1.82)
	95%	2.45 (2.18-2.69)	2.51 (2.17-2.89)
CS	5%	1.98 (1.39-2.59)	2.11 (1.76-2.55)
	Median	3.47 (3.08-3.85)	3.35 (2.97-3.76)
	95%	4.96 (4.34-5.54)	5.31 (4.39-6.35)

References

- Allen MR, Ingram WJ (2002) Constraints on future changes of climate and the hydrologic cycle. *Nature* 419: 224-231
- Cubasch U, Meehl GA, Boer GJ, Stouffer RJ, Dix M, Noda A, Senior CA, Raper S, Yap KS (2001) Projections of future climate change. In: Houghton JT et al (eds) *Climate Change 2001*. Cambridge University Press, pp 525-582
- Meehl GA, Boer GJ, Covey C, Latif M, Stouffer RJ (2000): The Coupled Model Intercomparison Project (CMIP). *Bull Am Meteorol Soc* 81: 313-318
- Nozawa T, Emori S, Takemura T, Nakajima T, Numaguti A, Abe-Ouchi A, Kimoto M (2000) Coupled ocean-atmosphere model experiments of future climate change based on IPCC SRES scenarios. Preprints, 11th Symposium on Global Change Studies, 9-14 January 2000, Long Beach, USA, 352-355

Annex 2: Extended Abstracts

Interpretation of Model Results that Show Changes in the Effective Climate Sensitivity with Time

Or: Is the climate sensitivity dead?

S. C. B. Raper

Alfred Wegener Institute, Germany

The climate sensitivity is defined as the equilibrium warming for a doubling of the concentration of CO₂, expressed as

$$\Delta T_{2x} = \Delta Q_{2x} / \lambda,$$

where ΔQ_{2x} is the doubled CO₂ forcing (Wm²) and λ is the feedback parameter (Wm⁻²/K). The climate sensitivity was a very useful concept in comparison of General Circulation Models for the classic experiment with doubled CO₂ forcing and a mixed layer ocean. But these mixed layer ocean models are not the ones used for transient simulations and the feedbacks/sensitivity may be different (see Boer and Yu 2003a). The concept of climate sensitivity was extended to transient simulations with time-varying forcing using coupled Atmosphere Ocean General Circulation Models (AOGCMs) by Murphy (1995). Murphy defined the ‘effective climate sensitivity’, denoted $\Delta T_{2x}^{\text{eff}}$, as the equilibrium response to a doubling of CO₂ which would occur if the AOGCM was run to equilibrium with feedback strengths, realised at a particular time, held fixed. The equation for the feedback parameter then includes a heat flux below the atmosphere denoted ΔF ,

$$\lambda = (\Delta Q - \Delta F) / \Delta T$$

and the effective climate sensitivity can be expressed as

$$\Delta T_{2x}^{\text{eff}} = \Delta Q_{2x} \cdot \Delta T / (\Delta Q - \Delta F).$$

The effective climate sensitivity is thus a hypothetical concept even within the realm of climate modelling. However, changes in the strengths of the climate feedbacks that accompany changes in climate and the rate of climate forcing are of great interest and importance for climate prediction and the effective climate sensitivity may be a useful way to quantify them. Following Murphy, several publications have shown that the effective climate sensitivity in transient runs with AOGCMs can vary with time but their interpretation is obscure (Murphy 1995, Watterson 2000, Senior and Mitchell, 2000, Raper et al. 2001).

The interpretation of varying effective climate sensitivity is not straightforward. This is in part due to the fact that it is a global derivative of feedbacks that occur on a local scale (see the detailed studies of Boer and Yu 2003b, 2003c). As a result, the effective climate sensitivity generally varies even in simulations with a simple model such as MAGICC (Figure 1). In the case

of the MAGICC model tuned to the AOGCM PCM and used in the TAR, for a step forcing of 3.6 Wm⁻², $\Delta T_{2x}^{\text{eff}}$ decreases with time (Figure 1d, X 0.5 Wm⁻²/K). It is only in the special case where the land/ocean exchange coefficient $X \approx 16$ Wm⁻²/K that the effective climate sensitivity is constant. In this case, a linear fit to a plot of global temperature versus flux into the surface crosses the zero temperature line at 3.6 Wm⁻¹, the value of the forcing (see Figure 1c, after Gregory et al. 2004), in the other two cases shown it does not. When $X = 16$ Wm⁻²/K the ratio of the land/ocean temperature change is a constant and the differing feedbacks of land and ocean ($\lambda_L - 4.9$, $\lambda_O - 6.6$ Wm⁻²/K) are given equal weight with time. When $X < 16$ Wm⁻²/K the land/ocean temperature change ratio decreases with time (see Figure 1b), when $X > 16$ Wm⁻²/K the land/ocean temperature change ratio increases with time, resulting in the changes in $\Delta T_{2x}^{\text{eff}}$ shown in Figure 1d. The former appears to be similar to the behaviour of the CCCma model in Boer and Yu, 2003c, whereas the latter is similar to the behaviour of the HadCM2 2xCO₂ experiment (Raper et al. 2001),

Figure 2 is analogous to Figure 1 but shows AOGCM results for the HadCM2 2xCO₂ experiment mentioned above. Notice the change in the relationship between the four surfaces box temperatures and the global mean temperature (Figure 2b) before and after forcing stabilization (black vertical line at 1.6 K). To train the eye, Figure 2c after Gregory et al (2004), includes the expected slope of the results after stabilization if the climate sensitivity were a constant of 2.6 K. Clearly the slope is much less than this and indicates an intercept for zero temperature change well below the forcing value of 3.471 Wm⁻². This is similar to the situation in Figure 1c for $X = 100$. Consistent with the findings of Figures 2b and c, the effective climate sensitivity for the HadCM2 2xCO₂ experiment shows an increase in time (similar to that for $X = 100$ in Figure 1d). (See Raper et al. 2001 for other forcings.)

Finally, as an illustration (not to be confused with a diagnosis), Figure 2d demonstrates that an increase in the effective climate sensitivity can be compatible with constant but different feedback parameters over the land and ocean, given the land and ocean temperature changes from the HadCM2 2xCO₂ experiment.

How, then, to define the climate sensitivity for a coupled AOGCM? Given that it is problematic to define and calculate the climate sensitivity for an AOGCM and the confusion that this quantity so often engenders in the non-scientist anyway – mistaking it for the projected range of changes for 2100 for example – it maybe that it is time to stop reporting about it in the Summary for Policymakers. A more transparent and relevant quantity for model comparison was introduced in the TAR: namely the Transient Climate Response (Cubasch et al., 2001).

References

Boer, G.J. and B. Yu, 2003a: Dynamical aspects of climate sensitivity. *Geophys. Res. Lett.* 30, 1135 (doi:10.1029/2002GL016549).
 Boer, G.J. and B. Yu, 2003b: Climate sensitivity and response. *Clim. Dyn.*, 20, 415-429.
 Boer, G.J. and B. Yu, 2003c: Climate sensitivity and climate state. *Clim. Dyn.*, 21, 167-176.
 Cubasch, and co-authors, 2001: Projections for future climate change. *Climate Change 2001: The*

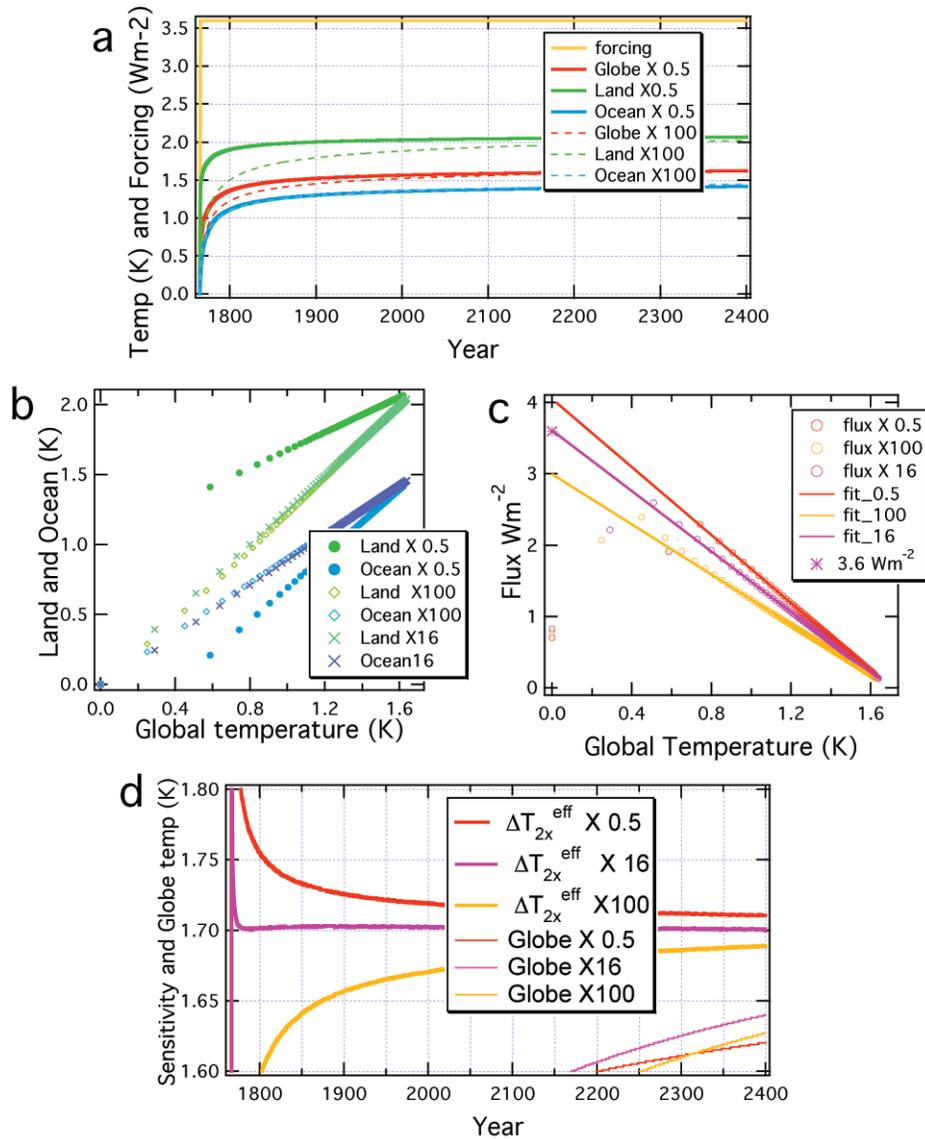


Figure 1. Illustration of the effect of the land/ocean heat exchange coefficient (X , Wm^{-2}/K) on the effective climate sensitivity using MAGICC tuned to PCM.

Scientific Basis, J. T. Houghton, et al., Eds., Cambridge University Press, 525–582.
 Gregory, J.M., W.J. Ingram, M.A. Palmer, G.S. Jones, P.A. Stott, R.B. Thorpe, J.A. Lowe, T.C. Johns, and K.D. Williams, 2004: A new method for diagnosing radiative forcing and climate sensitivity. *Geophys. Res. Lett.* 31, L03205, (doi: 10.1029/2003GL018747)
 Raper, S.C.B., J.M. Gregory and T.J. Osborn, 2001a: Use of an upwelling diffusion energy balance climate

model to simulate and diagnose AOGCM results. *Clim. Dyn.*, 17, 601-613.
 Senior, C.A. and J.F.B. Mitchell, 2000: The time-dependence of climate sensitivity. *Geophys. Res. Lett.*, 27, 2685-2688.
 Watterson, I.G. 2000: Interpretation of simulated global warming using a simple model. *J. Climate*, 13, 202-215.

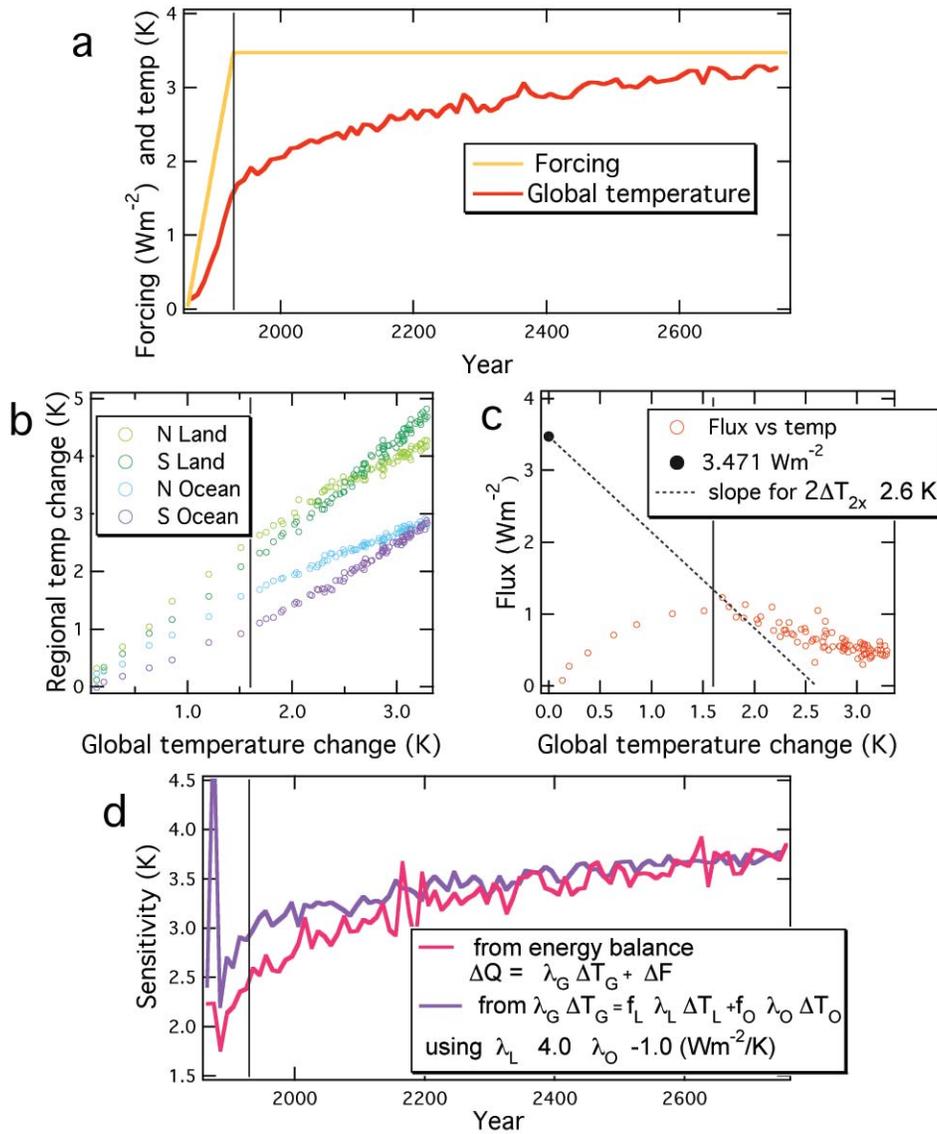


Figure 2. Illustration of how changing land/ocean heat flux could result in the time varying effective climate sensitivity found in a long HadCM2 integration.

Annex 2: Extended Abstracts

Use of Observations from the Mt Pinatubo Eruption to Estimate Climate Sensitivity

Alan Robock¹ and Piers Forster²

¹ Department of Environmental Sciences, Rutgers University, New Brunswick, New Jersey, USA

² NOAA Aeronomy Laboratory, Boulder, Colorado, USA and also at Department of Meteorology, University of Reading, Reading, UK

The June 15, 1991 Mount Pinatubo eruption was a large but relatively short-lived shock to the Earth's atmosphere. It thus provided an excellent opportunity to study the workings of the climate system and to test climate models, including the dynamical response to volcanically-produced temperature gradients in the lower stratosphere from aerosol heating and volcanic ozone depletion, as well as aspects of the carbon cycle and the impacts of climate change. Here we focus on using this natural experiment to test the sensitivity of the climate system to external radiative forcing. A comprehensive survey of the effects of volcanic eruptions on climate was presented by Robock [2000], updated by Robock [2002a, 2002b, 2003a, 2003b, 2004].

Most causes of climate change are gradual shifts in atmospheric composition or land surface characteristics. Volcanic eruptions, however, can produce a very large,

but short-lived, perturbation to the Earth's radiative balance. While we cannot use these perturbations to test long-term processes, such as changes in the thermohaline circulation, we can take advantage of them to examine some short time-scale feedback processes and impacts. Here we discuss two such aspects, the water vapor feedback as investigated in global climate models by Soden *et al.* [2002] and Forster and Collins [2004], and the global temperature response and recovery from individual volcanic eruptions, as discussed by Wigley *et al.* [2004].

Figure 1 shows the lower troposphere temperature anomalies for the Northern Hemisphere summer of 1992, one year after the Pinatubo eruption. Virtually the entire planet was cooler than normal, as expected. But the amount of cooling depended on the sensitivity of the climate system to radiative perturbations, and the

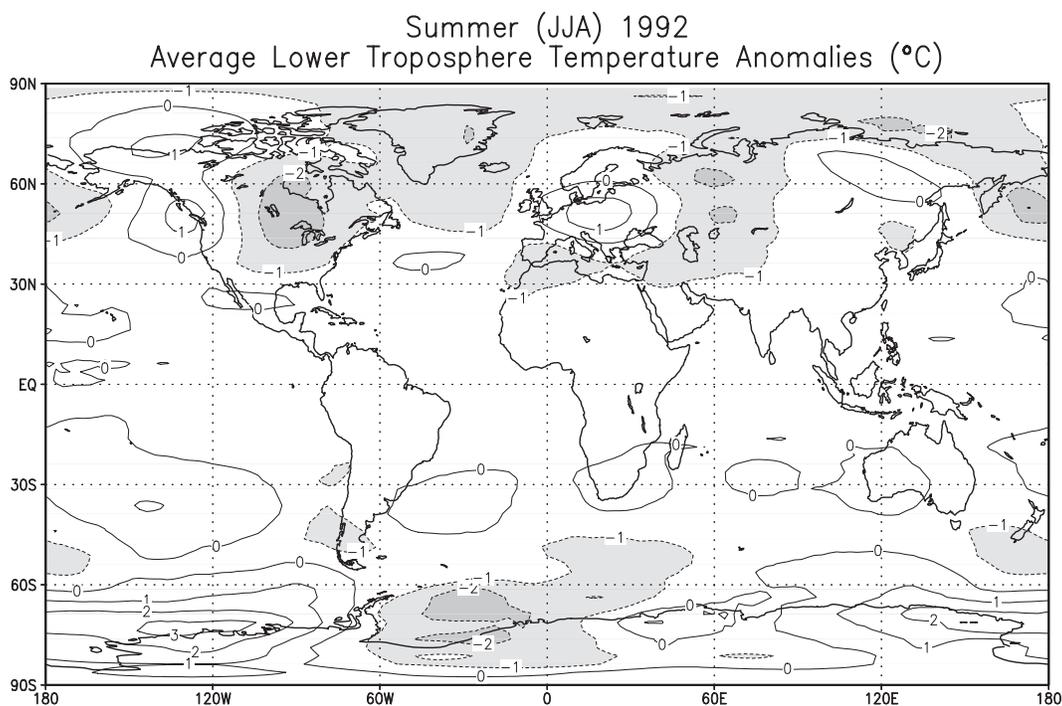


Figure 1. Summer (JJA) 1992 lower tropospheric temperature anomalies (with the non-volcanic period of 1984-1990 used to calculate the mean) following the 1991 Mt. Pinatubo eruption. Data from Microwave Sounding Unit Channel 2R [Spencer *et al.*, 1990], updated courtesy of J. Christy and now called Channel 2LT. Anomalies less than -1°C are shaded. (Figure 4 of Robock [2003b].)

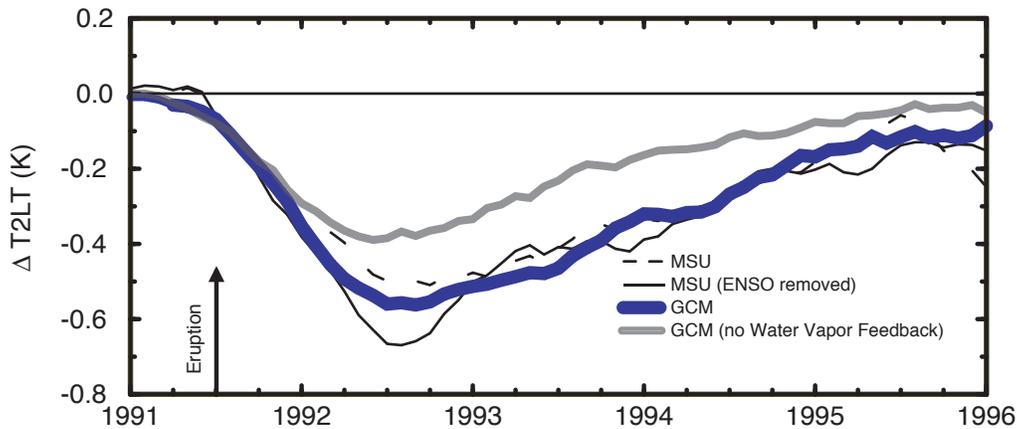


Figure 2. Observations (MSU – Microwave Sounding Unit [Spencer *et al.*, 1990] and MSU (ENSO removed) – with the effects of sea surface temperatures removed) and climate model simulations (GCM – general circulation model). The simulation was only successful with the positive water vapor feedback. (Figure 5 of Robock [2003b], from Figure 4 from Soden *et al.* [2002].)

strength of the most important positive feedback in the climate system, the water vapor-greenhouse feedback [Schneider and Dickinson, 1974]. As the planet cools, the amount of water vapor in the atmosphere goes down, reducing the greenhouse effect, and amplifying the cooling. (Of course, this positive feedback also works for warming.) The timing and amplitude of future global warming depend on this sensitivity of the climate system.

Soden *et al.* [2002] used the global cooling and drying of the atmosphere that was observed after the eruption of Mount Pinatubo to test model predictions of the climate feedback from water vapor, using the Geophysical Fluid Dynamics Laboratory R30 climate model. By comparing model simulations with and without water vapor feedback, they demonstrated the importance of atmospheric drying in amplifying the temperature change and showed that, without the strong positive feedback from water vapor, their atmospheric general circulation model was unable to reproduce the observed cooling (Figure 2). These results provide quantitative evidence of the reliability of water vapor feedback in current climate models, which is crucial to their use for global warming projections.

The Soden *et al.* [2002] simulations were done with fixed cloudiness. The R30 model with interactive cloudiness has a ΔT_{2x} sensitivity of 3.4 K, and the sensitivity of the fixed cloudiness model would be less than this, but it has not been determined. To reproduce the observed temperature record after the eruption of Mount Pinatubo, the model requires a strong positive feedback, equivalent in magnitude to that predicted for water vapor. Although it is possible that other processes, such as clouds, could act in place of water vapor to

provide the strong positive feedback necessary to amplify the cooling, the observational evidence clearly indicates a reduction in water vapor that is consistent with the model predictions.

Forster and Collins [2004] repeated parts of the Soden *et al.* [2002] study with the Hadley Centre coupled climate model (HADCM3). They analyzed 13 ensemble integrations of the climate response to the Mt. Pinatubo eruption and again compared these data to observations. They extended the comparison into the upper troposphere and also examined latitude-height patterns of the water vapor feedback. A water vapor feedback parameter of $-1.6 \text{ Wm}^{-2}\text{K}^{-1}$ was found in the observations, which agreed with that found for the model ensembles. Uncertainty, principally from natural climate variations that contaminated the volcanic cooling, placed the feedback parameter between -0.9 and $-2.5 \text{ Wm}^{-2}\text{K}^{-1}$. However, the study sounded a note of caution, as the latitude-height pattern of observed water vapor response following the eruption of Mt Pinatubo differed from that found in any ensemble integration of the model.

The experiments described above examine an important climate feedback mechanism using the best known large volcanic eruption. However, there have been several large eruptions in the recent past, and the sensitivity of the climate response to these eruptions, including the maximum cooling and the time taken for recovery can be used to determine the sensitivity of the climate system. Wigley *et al.* [2004] recently used the NCAR PCM climate model runs to evaluate an upwelling-diffusion energy balance model, and then examined the energy balance model response to the five largest eruptions of the past century, Santa Maria (1902),

Novarupta (Katmai) (1912), Agung (1963), El Chichón (1982), and Pinatubo (1991). The maximum cooling for any given eruption was shown to depend approximately on the climate sensitivity raised to power 0.37. After the maximum cooling, for low-latitude eruptions, the temperature relaxed back towards the initial state with an e-folding time of 29-43 months for ΔT_{2x} sensitivities of 1-4 K. The cooling associated with Pinatubo appear required a sensitivity above the IPCC lower bound of 1.5°C, and none of the observed eruption responses ruled out a sensitivity above 4.5 K. They also showed that the lower sensitivity estimates of Lindzen and Giannitis [1998], using a similar technique, were flawed because their climate model was inaccurate, and could not reproduce the behavior of more sophisticated general circulation models.

These results all support the current estimates of the sensitivity of the climate system to doubling CO₂. However, because the forcing from volcanic eruptions is not precisely known, because the climate response cannot be uniquely associated with volcanic forcing because of El Niño and stochastic variations, and because volcanic eruptions do not provide a perfect proxy for global warming, because the nature of the external radiative forcing obviously differs between the two, there are limitations to how precisely volcanic eruptions can be used to determine the climate sensitivity.

References

- Forster, Piers M. de F., and Matthew Collins, Quantifying the water vapour feedback associated with post-Pinatubo global cooling, *Climate Dyn.*, DOI: 10.1007/s00382-004-0431-z, 2004.
- Lindzen, R. S. and C. Giannitis, On the climatic implications of volcanic cooling. *J. Geophys. Res.*, 103, 5929-5941, 1998.
- Robock, A., Volcanic eruptions and climate, *Rev. Geophys.*, 38, 191-219, 2000.
- Robock, A., Pinatubo eruption: The climatic aftermath, *Science*, 295, 1242-1244, 2002a.
- Robock, A., Blowin' in the wind: Research priorities for climate effects of volcanic eruptions. *EOS*, 83, 472, 2002b.
- Robock, A., Volcanoes: Role in climate, in *Encyclopedia of Atmospheric Sciences*, J. Holton, J. A. Curry, and J. Pyle, Eds., (Academic Press, London), 10.1006/rwas.2002.0169, 2494-2500, 2003a.
- Robock, Alan, Introduction: Mount Pinatubo as a test of climate feedback mechanisms, In *Volcanism and the Earth's Atmosphere*, Alan Robock and Clive Oppenheimer, Eds., Geophysical Monograph 139 (American Geophysical Union, Washington, DC), 1-8, 2003b.
- Robock, Alan, Climatic impact of volcanic emissions, *State of the Planet*, Steve Sparks and Chris Hawkesworth, Eds., Geophysical Monograph, (American Geophysical Union, Washington, DC), in press, 2004.
- Schneider, S. H., and R. E. Dickinson, Climate modeling, *Rev. Geophys. Space Phys.*, 12, 447-493, 1974.
- Soden, B. J., R. T. Wetherald, G. L. Stenchikov, and A. Robock, Global cooling following the eruption of Mt. Pinatubo: A test of climate feedback by water vapor, *Science*, 296, 727-730, 2002.
- Spencer, R. W., J. R. Christy, and N. C. Grody, Global atmospheric temperature monitoring with satellite microwave measurements: Method and results 1979-1984, *J. Climate*, 3, 1111-1128, 1990.
- Wigley, T. M. L., C. M. Ammann, B. D. Santer, and S. C. B. Raper, The effect of climate sensitivity on the response to volcanic forcing, submitted to *J. Climate*, 2004.

METRIC: A Comparison of Climate Sensitivity from Different Forcings

Robert Sausen

*DLR-Institut für Physik der Atmosphäre, Oberpfaffenhofen, Germany***Introduction**

When comparing radiative forcings (RF) from different perturbations (e.g., from CO₂, CH₄ and CFCs) or when using CO₂ equivalents in the Kyoto Protocol we implicitly assume that the climate sensitivity parameter λ is constant. Under this assumption we can calculate the expected equilibrium response of the global mean near surface temperature ΔT_s to a given radiative forcing RF by a simple linear equation:

$$\Delta T_s = \lambda RF.$$

For many perturbations of the climate system like perturbations of well-mixed greenhouse gases or changes of the solar constant, the assumption is correct within one single climate model. However, the climate sensitivity parameter λ is sensitive to the model selected, ranging from 0.5 K/Wm⁻² to 1.3 K/Wm⁻² for a wide range of climate models. This variation in λ is believed to reflect differences in model parameterisations.

Many recent climate simulations show that the climate sensitivity parameter cannot be considered as a constant in the case of inhomogeneously (vertically or/and geographically) distributed climate forcing agents, not even within one climate model. This is the case for, e.g., ozone perturbations in selected altitudes (Hansen et al., 1997; see also IPCC, 1999, their Fig. 6-10) or for aircraft-induced ozone perturbations (Ponater et al., 1999; Stuber et al., 2001a). This finding was systematically explored in the EU project METRIC.

Idealised perturbations

In a first set of numerical experiments, the climate impact (in terms of surface temperature change) of idealised geographically inhomogeneous perturbations of radiatively active species was studied with two comprehensive GCMs, ECHAM4 run by the DLR-Institut für Physik der Atmosphäre, Germany, and LMDz run by Laboratoire de Météorologie Dynamique, CNRS, France, and one intermediate GCM (UREAD) run by the University of Reading, United Kingdom (Sausen et al., 2002; Joshi et al., 2003). Each model includes a thermodynamic model of the oceanic mixed layer and the sea-ice.

The following perturbations, each scaled to have the same global mean radiative forcing of 1 Wm⁻², were considered:

- SG* solar constant increased globally
- ST* solar constant increased in the tropics, i.e. equatorwards of 30°

- SE* solar constant increased in the extra-tropics, i.e. polewards of 30°

- CG* CO₂ added globally, vertically homogeneous

- CT* CO₂ added in the tropics, vertically homogeneous

- CE* CO₂ added in the extra-tropics, vertically homogeneous

- CN* CO₂ added northwards of 30°N, vertically homogeneous

- OG-UT* ozone added globally in the upper troposphere

- OT-UT* ozone added equatorwards of 30° in the upper troposphere

- OE-UT* ozone added polewards of 30° in the upper troposphere

- ON-UT* ozone added northwards of 30°N in the upper troposphere

- OG-LS* ozone added globally in the lower stratosphere

- ON-LS* ozone added northwards of 30°N in the lower stratosphere

- ON-LT* ozone added northwards of 30°N in the planetary boundary layer

Perturbations in the solar constant and CO₂ concentration mainly impact the short-wave and long-wave radiation, respectively, whereas O₃ changes are of importance for both, the short-wave and long-wave radiation. Fig. 1 shows the resulting climate sensitivity parameters for the various perturbations. Obviously, there is a high variability from model to model and also from perturbation to perturbation.

Introducing the normalised climate sensitivity parameter

$$r = \lambda / \lambda_{\text{CO}_2},$$

i.e., dividing any climate sensitivity parameter by the climate sensitivity parameter for CO₂ of the same model, removes some of the model dependence.

Now similarities in the response to the perturbations become apparent (Fig. 2). The following general results are robust across the models: The tropical and extra-tropical perturbations result in smaller or larger temperature changes, respectively. The response to ozone perturbations in the upper troposphere tends to be smaller than for CO₂, while the response to the ozone increase in the lower stratosphere is the largest for each model.

In the case of the ozone increase in the lower stratosphere, the particular high climate sensitivity parameter appears to be a consequence of enhanced stratospheric water vapour (Stuber et al., 2001b, 2004).

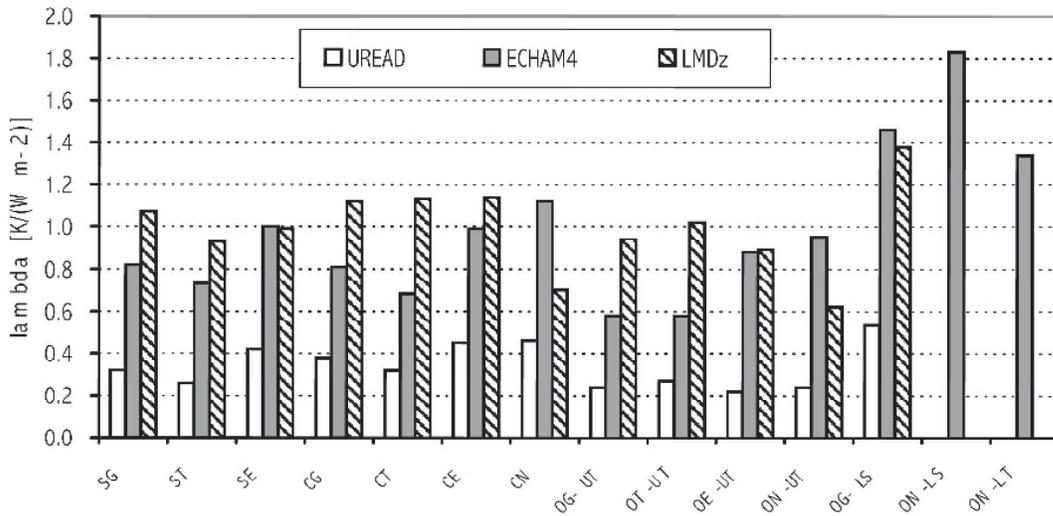


Figure 1. Climate sensitivity parameters λ for various perturbations as simulated with the UREAD, ECHAM4 and LMDz models (Sausen et al., 2002; Joshi et al., 2003).

In the simulations with the ECHAM4 model a 1 Wm^{-2} homogeneous CO_2 increase (CG) results in an increase of the stratospheric water vapour concentrations of about 20%, whereas the radiatively equivalent ozone perturbation in the lower stratosphere (OG-LS) leads to an increase in stratospheric water vapour by more than 200%. It appears that in the OG-LS experiment the strongest primary warming is close to the tropical tropopause. Consequently, the cold point temperature grows, and more water vapour can enter the stratosphere (Stuber et al., 2004). This starts a positive feedback loop.

The largest normalised climate sensitivity parameter r is found in the ON-LS, i.e., if the ozone perturbation is in the lower stratosphere of the northern

extra-tropics. Here the sea-ice-albedo feedback is also important (Stuber et al., 2004).

“Realistic” perturbations

In a second set of experiment more realistic perturbations were studied (Berntsen et al., 2004). First, the ozone changes resulting from additional emissions of CO (40 Tg/yr) and NOx (1 Tg(N)/yr in Europe and South-East Asia) were calculated with two (tropospheric) Chemical Transport Models (CTMs), the UiO (CTM2 of University of Oslo, run by CICERO, Norway) and the LMDzINCA (Laboratoire des Sciences du Climat et de l’Environnement, France) models. Later the resulting ozone perturbations were used as forcings of two GCMs

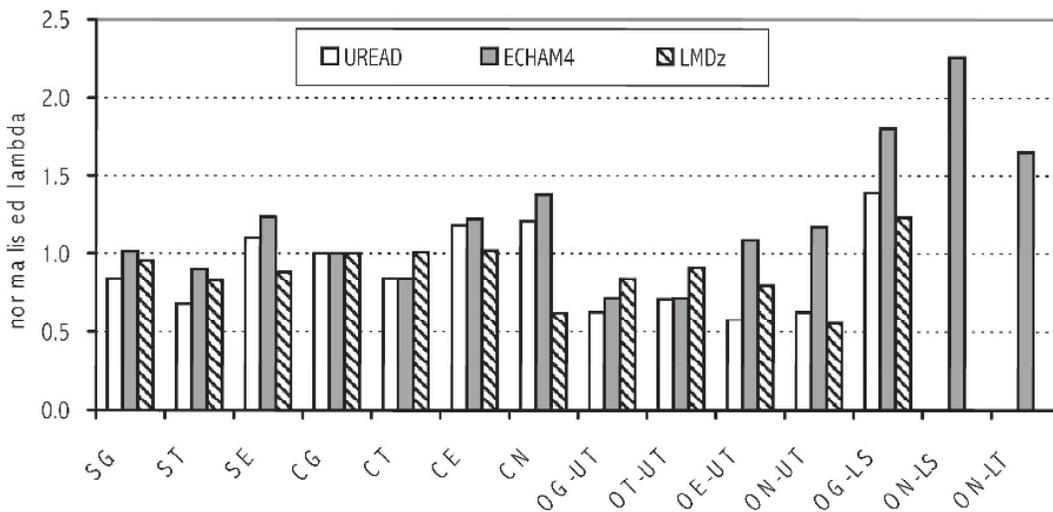


Figure 2. Normalised climate sensitivity parameters r for various perturbations as simulated with the UREAD, ECHAM4 and LMDz models (Sausen et al., 2002; Joshi et al., 2003).

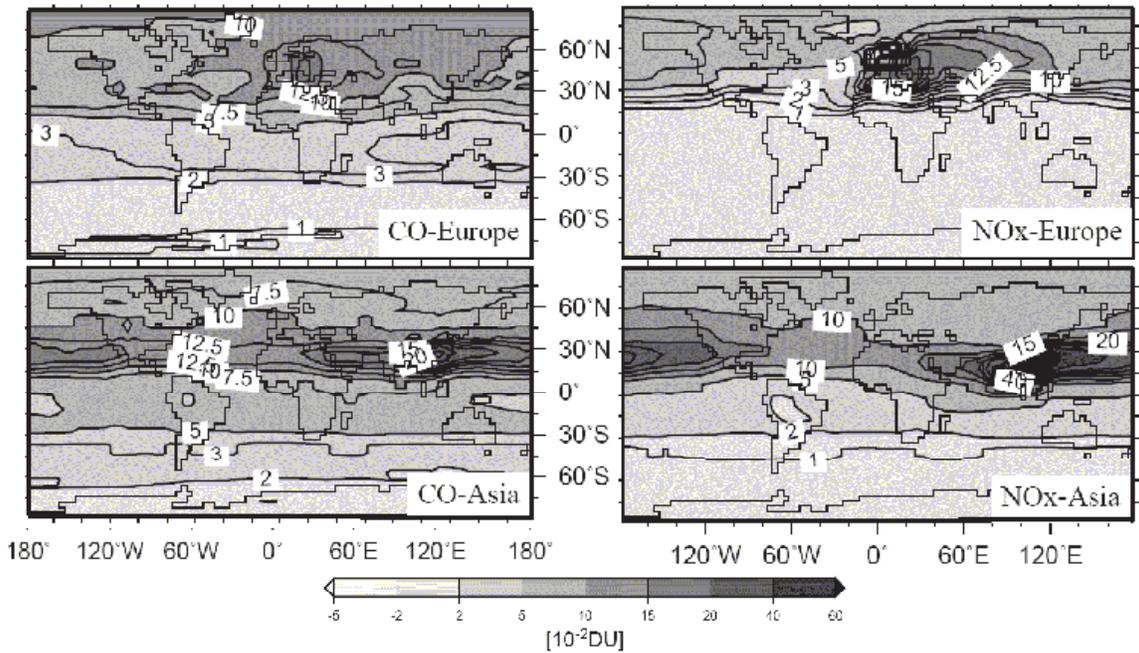


Figure 3. Annual mean change of the tropospheric ozone column due to CO and NOx emissions in Europe and South-East Asia as simulated by the UiO model (Berntsen et al., 2004).

(ECHAM4 and UREAD), after scaling to give a 1 Wm^{-2} radiative forcing.

The simulated changes in the ozone concentrations mainly depend on the location of the emissions. Model-to-model differences are of minor importance for the CTMs applied. Due to the shorter atmospheric life time of NOx relative to CO, the NOx-induced ozone perturbations are more localised than the CO-induced changes. Fig. 3 shows annual mean changes tropospheric ozone columns for the four cases as simulated by the UiO model.

The geographical distributions of the radiative forcings resulting from the NOx-induced ozone perturbations are quite different for emissions in Europe and in South-East Asia (Fig. 4). The higher values are found for emissions in South-East Asia as was the case for the ozone changes (Fig. 3).

The CO and NOx emissions also impact on the chemical life time of methane: the CO emissions result in an increase of the life time whereas the NOx emissions decrease the life time. Consequently, the abundance of the greenhouse gas methane increases

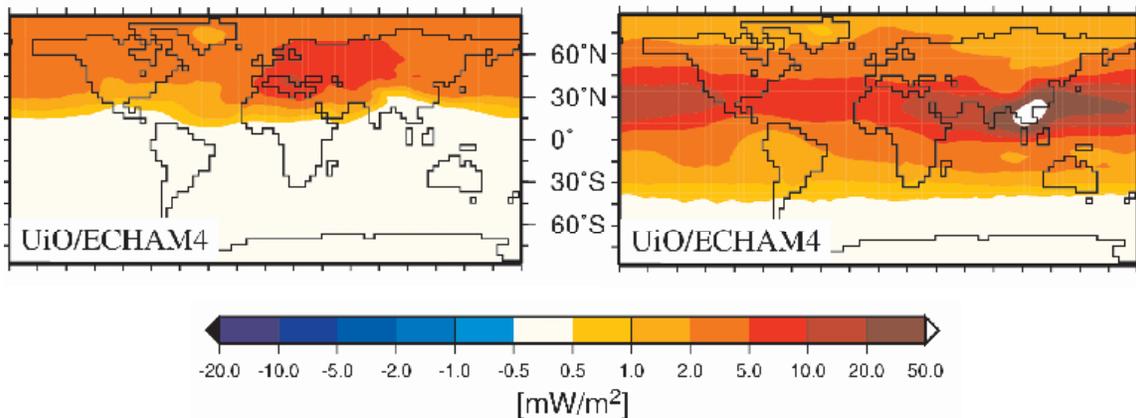


Figure 4. Annual mean radiative forcings from ozone changes due to NOx emissions in Europe (left panel) and South-East Asia (right panel) as calculated by ECHAM4 based on the ozone from the UiO model (Berntsen et al., 2004).

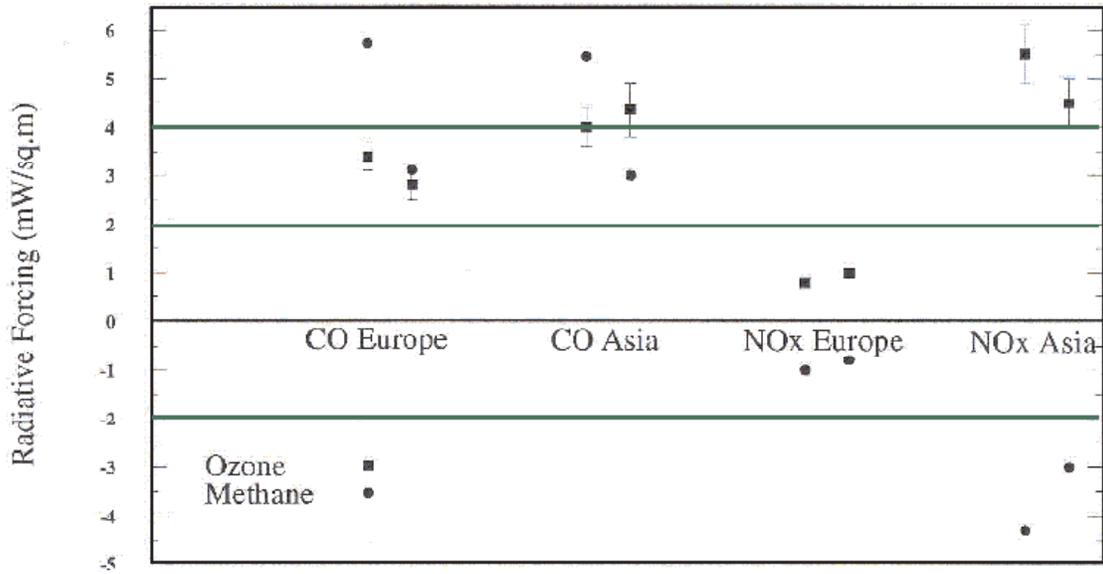


Figure 5. Annual mean radiative forcings from ozone and methane changes due to CO and NOx emissions in Europe and Southeast Asia (Berntsen et al., 2004).

and decreases for CO and NOx emissions, respectively. Fig. 5 summarises the radiative forcings arising from CO and NOx emissions in Europe and South-East Asia.

The “realistic” ozone perturbations were scaled to give a radiative forcing of 1 Wm⁻² and then inserted into the ECHAM4 and UREAD models. Despite the great differences in the geographical structures of the radiative forcings, the patterns of the resulting temperature change are more similar (Fig. 6).

As for the idealised perturbations, the climate sensitivity parameters differ strongly between ECHAM4 and UREAD, which is a consequence of the different climate sensitivities of the two models. Here we concentrate on the normalised climate sensitivity parameters as defined above (Fig. 7). While the ECHAM4 model tends to higher climate sensitivities for non-CO₂ emissions, the UREAD model shows the contrary results.

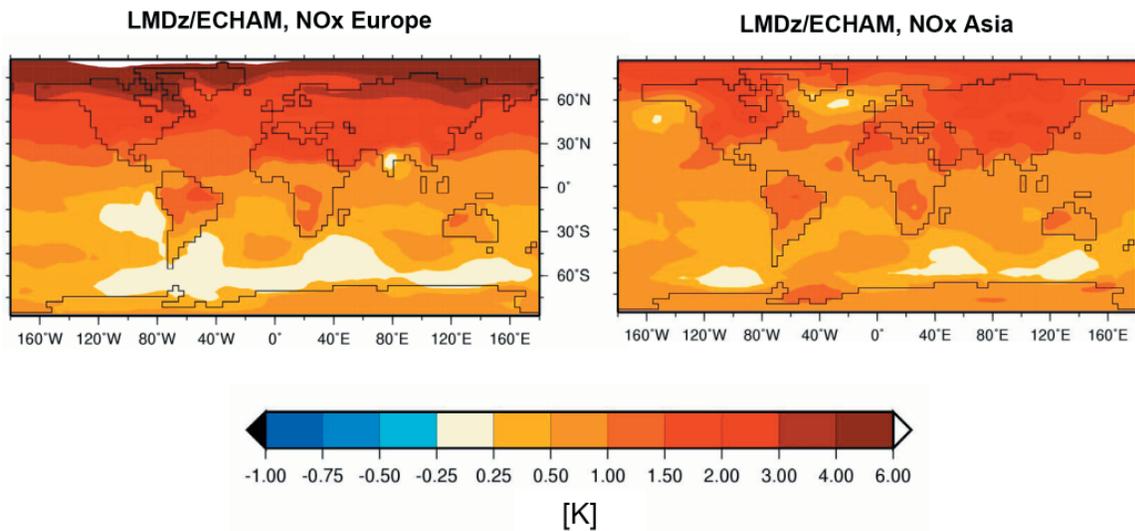


Figure 6. Annual mean surface temperature change from NOx-induced ozone perturbations for NOx emissions in Europe (left) and Southeast Asia (right) as simulated by ECHAM4 based on the ozone from the UiO model (Berntsen et al., 2004).

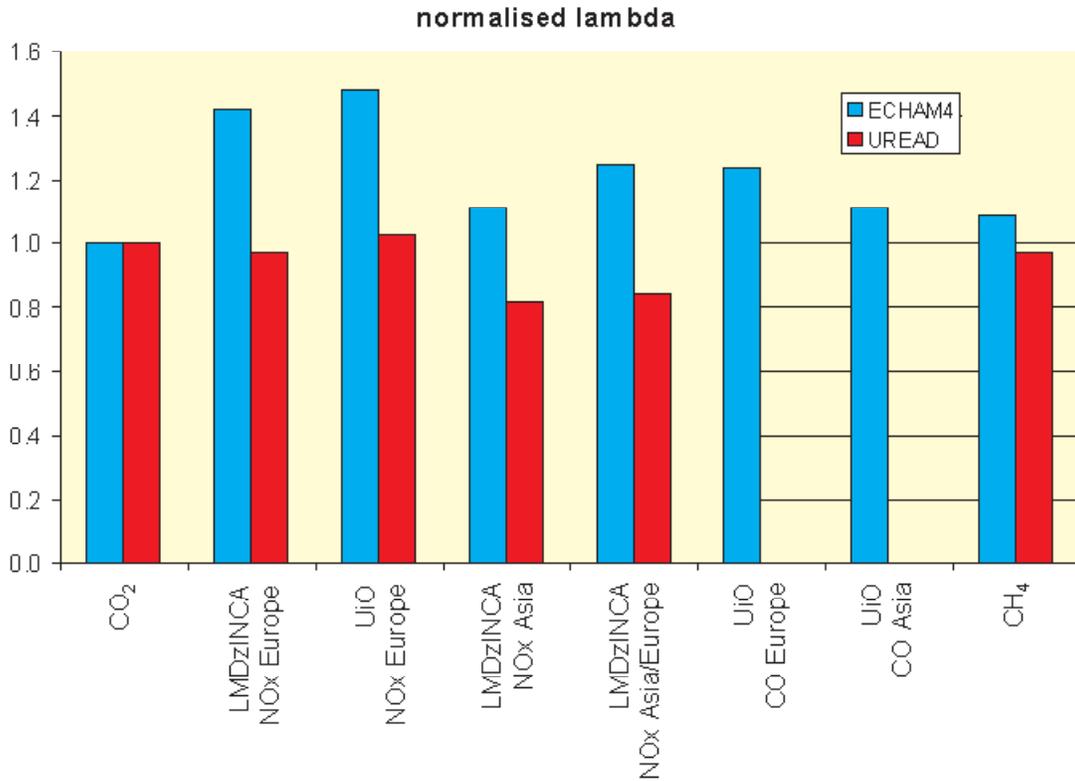


Figure 7. Normalised climate sensitivity parameters r for various “realistic” perturbations as simulated with the UREAD and ECHAM4 models (based on Bernsten et al., 2004).

Refined metrics

The global warming potential GWP of a species i , the central metric in the Kyoto Protocol, is defined as

$$GWP(H)_i = \frac{\int_0^H RF_i(t) dt}{\int_0^H RF_{CO_2}(t) dt} = \frac{\int_0^H a_i c_i(t) dt}{\int_0^H a_{CO_2} c_{CO_2}(t) dt},$$

where H is the time horizon (100 years in the Kyoto Protocol), RF_j the radiative forcing of species j , a_j and c_j the radiative forcing per unit and the concentration, respectively, of species j .

The effect of differences in climate response to inhomogeneously distributed forcing agents can be included in this metric by the following modification

$$GWP(H)^* = \frac{\int_0^H \lambda_i RF_i(t) dt}{\int_0^H \lambda_{CO_2} RF_{CO_2}(t) dt} = r_i \cdot GWP(H).$$

Further concepts for refined metrics are the modified Sustained Global Warming Potential $SGWP^*$ (Bernsten

et al., 2004) and the Global Temperature Potential GTP (Shine et al., 2004).

Conclusions

- The global mean radiative forcings of a short-lived species depend on the location of the emissions of the species or their precursors.
- The climate sensitivity parameter is dependent on the type of forcing agents and their geographical and vertical distributions.
- Not only the basic climate sensitivity (for homogeneous CO₂ perturbations) but also its variance among different forcings is model-dependent.
- Despite some model-to-model similarities we are not yet able to recommend reliable measures for introducing short-lived climate change agents in climate protection treaties, as we insufficiently understand model dependencies.
- A multi-model approach is necessary in order to better extract the model independent information.
- Potentially superior metrics have been worked out and tested.

References

- Berntsen, T.K., J.S. Fuglestedt, M.M. Joshi, K.P. Shine, N. Stuber, M. Ponater, R. Sausen, D.A. Hauglustaine, and L. Li, 2004:** Climate response to regional emissions of ozone precursors; sensitivities and warming potentials. Submitted to *Tellus*.
- Hansen, J., M. Sato and R. Ruedy, 1997:** Radiative forcing and climate response. *J. Geophys. Res.* 102, 6831-6864.
- IPCC, 1999:** Aviation and the global atmosphere - A special report of IPCC working groups I and III. Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge, UK and New York, NY, USA, 365 pp.
- Joshi, M., K. Shine, M. Ponater, N. Stuber, R. Sausen and L. Li, 2003:** A comparison of climate response to different radiative forcings in three general circulation models: towards an improved metric of climate change. *Climate Dyn.* 20, 843-854.
- Ponater, M., R. Sausen, B. Feneberg, and E. Roeckner, 1999:** Climate effect of ozone changes caused by present and future air traffic. *Climate Dyn.* 15, 631-642.
- Sausen, R., M. Ponater, N. Stuber, M. Joshi K. Shine and L. Li, 2002:** Climate response to inhomogeneously distributed forcing agents. In J. van Ham, A.P.M. Baede, R. Guicherit and J.G.F.M. Williams-Jacobse (eds.): *Non-CO2 Greenhouse Gases: Scientific Understanding, Control Options and Policy Aspects*. Millpress, Rotterdam, Netherlands, ISBN 90-77017-70-4, 377-381.
- Shine, K.P., J.S. Fuglestedt, K. Hailemariam and N. Stuber, 2004:** Alternatives to the global warming potential for computing impacts of emissions of greenhouse gases. *Climatic Change*, in press.
- Stuber, N., R. Sausen, and M. Ponater, 2001a:** Stratosphere adjusted radiative forcing calculations in a comprehensive climate model. *Theor. Appl. Climatol.* 68, 125-135.
- Stuber, N., M. Ponater, and R. Sausen, 2001b:** Is the climate sensitivity to ozone perturbations enhanced by stratospheric water vapor feedback? *Geophys. Res. Lett.* 28, 2887-2890.
- Stuber, N., M. Ponater, and R. Sausen, 2004:** Why radiative forcing might fail as a predictor of climate change. Submitted to *Climate Dyn.*

Climate Sensitivity Range Derived from Large Ensemble Simulations of Glacial Climate Constrained by Proxy Data

T. Schneider von Deimling, H. Held, A. Ganopolski, and S. Rahmstorf

Recent studies suggest that the uncertainty range of climate sensitivity (ΔT_{2x}) is only weakly constrained by the anthropogenic warming signal¹⁻³. In our work we constrain ΔT_{2x} by focusing on the large radiative forcing change between the preindustrial and the glacial climate, by requiring consistency of modelled glacial cooling with paleo-data. We performed a large ensemble of glacial simulations and accounted for a complete set of the main radiative forcings. Our approach yields effective constraints on ΔT_{2x} , resulting in a maximum range (5 to 95%) of 1.5 to 4.7°C – with a best guess between 2.1 and 3.6°C.

The signal of anthropogenic warming is too weak to effectively constrain ΔT_{2x} in that uncertainty in the radiative forcing over the industrial period, in the observational data and in the rate of ocean heat uptake is comparatively large. As an alternative, we use paleo-data from the Last Glacial Maximum (LGM, 21kys B.P.) to constrain ΔT_{2x} . This period represents the largest deviation from present climate in recent geologic history, manifested in a pronounced global cooling. The LGM climate state is of high interest not only because of the large temperature signal, but also because the main forcings are well known⁴, well-dated paleo data are available, and the cold climate state lasted long enough to be in near-equilibrium. Hence, it is promising to combine model simulations and paleo-data for the LGM to derive constraints on climate sensitivity.

We performed ensemble simulations using a climate model of intermediate complexity (CLIMBER-2)^{5,6}, consisting of a statistical-dynamical 2.5-dimensional atmosphere coupled to a multi-basin, zonally averaged ocean model without need for flux adjustments. In our ensemble runs we consider uncertainties in eleven model parameters that strongly affect the main model feedbacks determining ΔT_{2x} , namely cloud-, water vapour- and lapse rate feedback.

Using pre-industrial boundary conditions we performed 1,000 CLIMBER-2 runs in which we varied the eleven parameters simultaneously. The climate sensitivity for each of the 1,000 model versions is calculated by running the model for 3,500 years into an equilibrium climate state of 280 ppm CO_2 , and subsequently into a new equilibrium of 560 ppm for another 3,000 years. With the chosen parameter ensemble, climate sensitivity ΔT_{2x} ranged from 1.3 to

5.5°C. Most parameter combinations result in large discrepancies of simulated pre-industrial climate and observational data. Hence, we constrained the ensemble to models consistent with global pre-industrial climate characteristics. The consistency criteria lead to a reduction of the original ensemble size by about 90%, which demonstrates that the parameter choices not only strongly affect the temperature response to CO_2 but also the pre-industrial climate. However, the preindustrial data constraints hardly reduced the range of climate sensitivity (in the subset of 111 models with realistic pre-industrial climate, ΔT_{2x} ranged from 1.4 to 4.9°C).

To test whether paleo-data can be used to reduce the uncertainty in the range of ΔT_{2x} , we then ran the full ensemble of models for LGM boundary conditions. The main forcing changes between pre-industrial and LGM climate are accounted for in the simulation by lowered GHG concentrations (CO_2 , CH_4 , N_2O), existence of northern hemisphere ice sheets, increased atmospheric dust concentration, vegetation and insolation changes. It should be understood that we determine the climate sensitivity directly from the simulated temperature increase due to CO_2 doubling in each model version, and not from the temperature change between the LGM and pre-industrial period. Our results thus automatically account for differences in feedbacks and climate response between colder and warmer climates; we do not assume that the sensitivity to CO_2 changes is the same when going to LGM conditions as it is for CO_2 doubling.

Figure 1 shows the relation between the simulated warming due to a doubling of CO_2 and the magnitude of LGM cooling for tropical and high latitude regions. The strong correlation between climate sensitivity and LGM cooling is striking. This close link is the basis for our approach to constraining ΔT_{2x} .

As estimates of global mean LGM cooling are uncertain due to the lack of global paleodata coverage, we focus on specific regions for our model-data comparison. Particularly suited are the tropical ocean regions, where data from sediment cores are abundant. Reconstructed SSTs from paleo-data have been discussed controversially over the past decades, particularly the magnitude of tropical temperature response. Yet in recent years the inter-comparison of different reconstruction techniques has led to reject very low (smaller than -1.5°C) and high (larger than

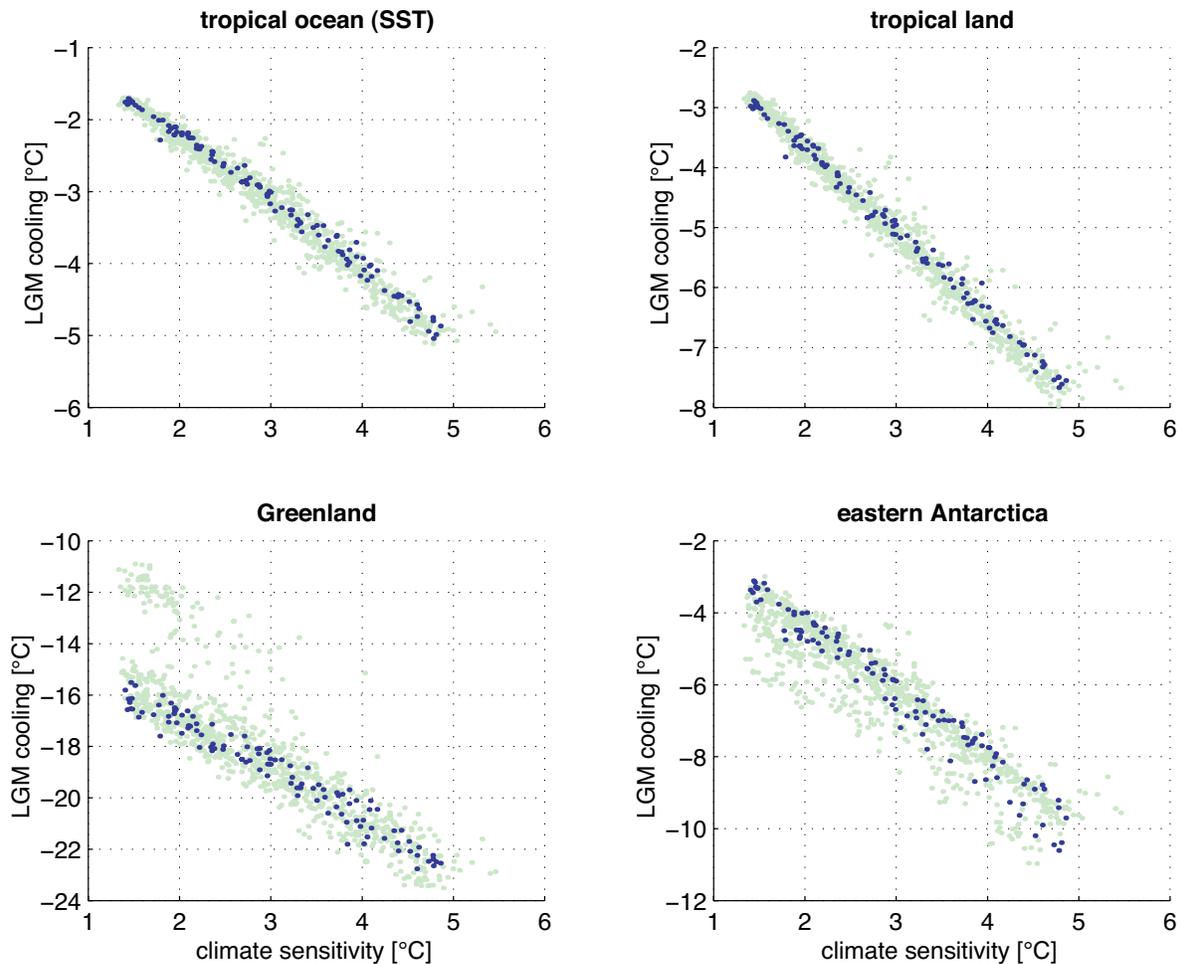


Figure 1. Dependence of LGM cooling (relative to the pre-industrial climate) on climate sensitivity for different regions. a, tropical (30°S-30°N) ocean, b, tropical land, c, Greenland, d, eastern Antarctica. Green points represent the entire ensemble (1,000 runs), blue points only runs consistent with pre-industrial data constraints (111 runs).

-4°C) tropical SST cooling estimates⁷. We represent the uncertainty of reconstructed ΔSST by considering three assumptions of tropical SST cooling as plausible: a) low (2-3°C), b) high (3-4°C) and c) best-guess (2.5-3°C)^{7,8}.

The climate sensitivity ranges (5 to 95 percentile), which are consistent with a specific tropical SST cooling (a-c), are illustrated in Fig. 2 (blue intervals). When considering a specific tropical SST cooling as most likely, the corresponding climate sensitivity range is notably smaller than that given in the IPCC TAR. Uncertainty in dust forcing is accounted for by performing a second LGM ensemble where the magnitude of dust forcing is reduced by 50%. This leads to a slight shift (by about 0.5°C) of the climate sensitivity intervals to larger values (light blue intervals). Our largest climate sensitivity range, covering all uncertainty in tropical cooling (cases a-c), dust forcing, and in model parameter choices, suggests a very

high probability for the climate sensitivity to lie within the range of 1.5 to 4.7°C.

Similar estimates of ΔT_{2x} can be derived using tropical land data, which are subject to larger data uncertainty. Paleo-data constraints from high latitude regions (Greenland, Antarctica) also give results consistent with those shown in Fig. 2. We thus conclude that very low (<1.5°C) and very high (>4.7°C) values of ΔT_{2x} are inconsistent with current understanding of LGM climate. The cooling during the LGM as mapped by paleo-climatic proxy data gives independent and strong support for midrange estimates of climate sensitivity. Reducing the uncertainty range in LGM data will automatically reduce the uncertainty range in climate sensitivity; e.g., if the current “best guess” value for LGM tropical cooling of 2.5-3°C is confirmed, our study implies that climate sensitivity lies in the range 2.1-3.6°C.

References

1. Forest, C. E., Stone, P. H., Sokolov, A. P., Allen, M. R. & Webster, M. D. Quantifying uncertainties in climate system properties with the use of recent climate observations. *Science* 295, 113-117 (2002).
2. Knutti, R., Stocker, T. F., Joos, F. & Plattner, G.-K. Constraints on radiative forcing and future climate change from observations and climate model ensembles. *Nature* 416, 719-723 (2002).
3. Andronova, N. & Schlesinger, M. E. Objective estimation of the probability distribution for climate sensitivity. *J. Geophys. Res.* 106, 22605-22612 (2001).
4. Hewitt, C. D., Stouffer, R. J., Broccoli, A. J., Mitchell, J. F. B. & Valdes, P. J. The effect of ocean dynamics in a coupled GCM simulation of the Last Glacial Maximum. *Climate Dynamics* 20, 203-218 (2003).
5. Petoukhov, V. et al. CLIMBER-2: a climate system model of intermediate complexity. Part I: model description and performance for present climate. *Climate Dynamics* 16, 1 (2000).
6. Ganopolski, A. et al. CLIMBER-2: a climate system model of intermediate complexity. Part II: model sensitivity. *Climate Dynamics* 17, 735 (2001).
7. Crowley, T. J. CLIMAP SSTs re-revisited. *Climate Dynamics* 16, 241-255 (2000).
8. Lea, D. W., Pak, D. K., Peterson, L. C. & Hughen, K. A. Synchronicity of tropical and high-latitude Atlantic temperatures over the last glacial termination. *Science* 301, 1361-1364 (2003).

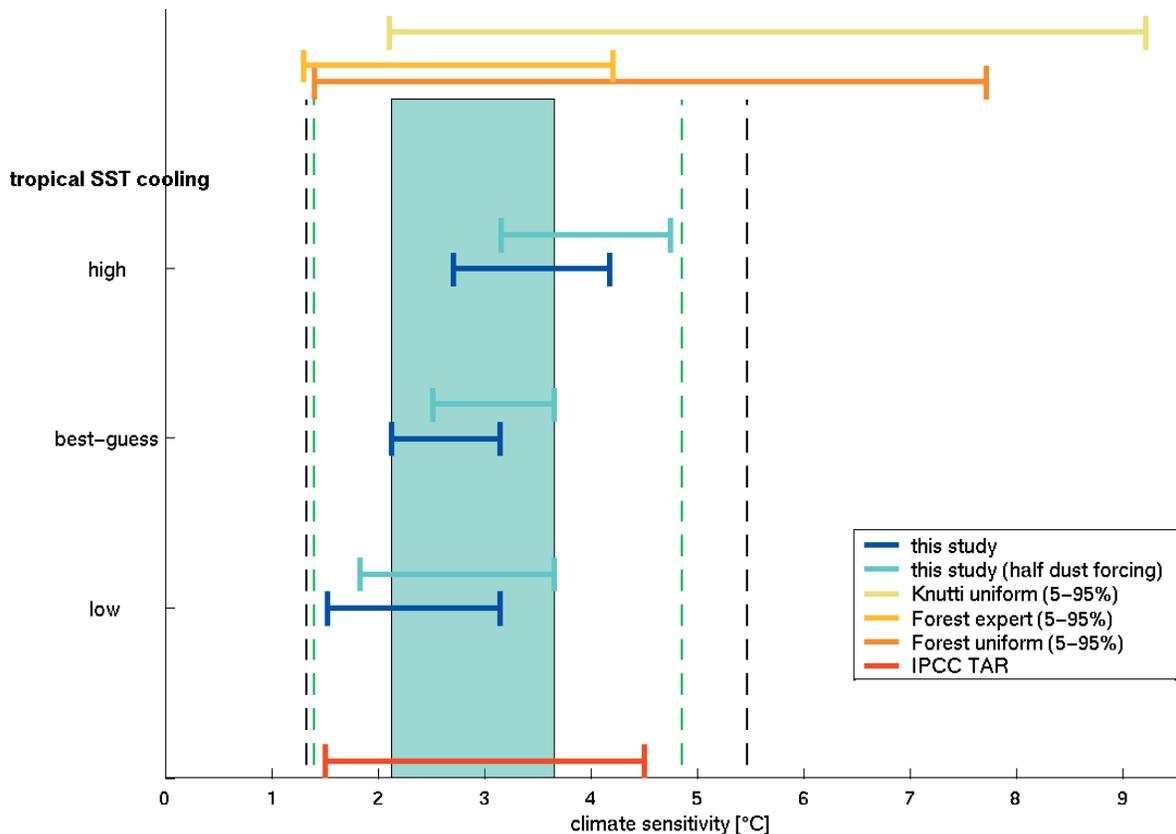


Figure 2. Climate sensitivity estimates. Dark blue intervals represent the ranges (5-95%) of ΔT_{2x} that are consistent with a low (2-3°C), best-guess (2.5-3°C), and high (3-4°C) tropical SST cooling. Light blue intervals show analogous ranges, which result from simulations where dust forcing is reduced by 50%. Vertical black (green) dashed lines represent the realised climate sensitivity range resulting from the complete (preindustrial data consistent) parameter ensemble. Yellow- and red coloured intervals show recent climate sensitivity estimates (see legend). The light blue box illustrates the range of ΔT_{2x} for our best-guess of tropical SST cooling when accounting for uncertainty in dust forcing.

Annex 2: Extended Abstracts

Report on the joint WCRP CFMIP/IPCC expert meeting on 'Climate Sensitivity and Feedbacks': Summary and Recommendations

Catherine Senior

Hadley Centre, Met Office, U.K.

Background

An expert meeting on 'Climate Sensitivity and feedbacks' was held at the Met Office in Exeter, U.K. from April 19-22nd, 2004. The meeting was jointly arranged by IPCC WG1 and the WCRP CFMIP (Cloud Feedback Inter-comparison project: <http://www.cfmip.net>), and acted both as a first meeting of CFMIP and as an expert meeting for IPCC WG1 to develop scientific questions relevant to the intergovernmental workshop on Climate Sensitivity, July 26th -29th in Paris. The Exeter meeting had as its main focus the role of clouds and cloud feedbacks in determining climate sensitivity. The meeting had 3 major sessions on 'Feedback Methods', 'Cloud: Validation and feedbacks' and 'CFMIP results and Plans'. In addition a number of key speakers gave presentations at a Royal Meteorological Society meeting on 'Climate Sensitivity and Feedbacks' held during the meeting. On the final day of the meeting, three breakout groups were asked to discuss topics that had arisen during the meeting. These topics were chosen to be issues on which we might be able to make recommendations to IPCC WG1 as ways forward to addressing some of the key scientific questions raised in the IPCC WG1 Workshop on Climate Sensitivity 'Scope and Aims' document. Here we summarise the topics of the breakout groups and the recommendations put forward

Breakout Groups (BOGs)

1. *BOG1: Climate Feedback Metrics.*

During the meeting, a number of speakers had talked about the possibility of constraining climate sensitivity using a basket of measures (or metrics) aimed at validating present day climate. BOG1 started by reviewing the utility of current experimental design (notably from the CFMIP project) for such a purpose and looked at what data would be needed to allow the necessary metric tests. The focus was particularly on clouds, although it was recognized that such an approach could be broadened as required. A number of climate model ensembles (the so-called QUMP ensemble at the Hadley Centre (Murphy et al, 2004), the climateprediction.net ensemble (Stainforth et al, 2004) and the forthcoming CFMIP ensemble are available

(or will be soon) and BOG1 propose a way forward for using such ensembles that aims to relate errors in model simulations (when compared to observations) to uncertainty in climate prediction. The ultimate goal being to reduce uncertainty in climate sensitivity through application of such carefully constructed metrics to a range of models.

2. *BOG2: CFMIP*

Those involved in CFMIP discussed the plans for the project with particular reference to IPCC WG1, and highlighted a number of key areas. A recommendation was made that the CFMIP data be used for research into developing climate metrics, notably related to clouds (stratocumulus and the link between upper tropospheric humidity and cirrus cloud were considered most urgent). Also it was recommended that the ISCCP-simulator (a software package that can be applied to model output in order to produce quantities that can be directly compared to quantities retrieved from the International Satellite Cloud Climatology Project) be included in IPCC AR4, AMIP and CMIP runs. It will be recommended to groups involved in CFMIP to diagnose instantaneous radiative forcing (as will be available for AR4 experiments) and to use partial radiative perturbations for analyzing cloud feedbacks in addition to the more common approach of looking at changes in cloud radiative forcing. A novel area of research for CFMIP is developing methodologies (e.g compositing and clustering) to evaluate aspects of clouds in models that can be shown to be important for cloud response to climate change. It is recommended that such techniques should be applied to a wider range of models, e.g. those in AMIP, CMIP and AR4. This could be best achieved if modeling groups who submit data to AR4 can be encouraged to also participate in CFMIP.

3. *BOG3: Climate sensitivity and response*

The third BOG discussed the relative roles and importance of transient and equilibrium climate sensitivity particularly to policy makers. Recommendations were that modeling groups should provide the temporal evolution of the effective (or transient) climate sensitivity to IPCC AR4 and should be encouraged to analyse the spatial and temporal

structure of the feedbacks under a range of IPCC forcings. There was discussion of the relationship between climate sensitivity, ocean heat uptake and temperature response and ways of constraining climate sensitivity and temperature response using the past behaviour of the climate system. One approach is the use of climate perturbations that are comparatively well known. A recommendation was made that IPCC AR4 coupled models include the simulation of the climate perturbation associated with the Mount Pinatubo eruption, paying particular attention to the nature and timescale of the simulated response, the role of ocean heat uptake and the bounds that can be placed upon climate sensitivity. Another approach is the use of observed trends (both in the instrumental record and in palaeoclimate reconstructions) to constrain climate sensitivity. A recommendation was made that studies of past forced climate variations should be pursued but will require carefully developed datasets of forcing and response. Finally constraining temperature evolution was discussed, as for small changes this is primarily dependent on forcing and ocean heat uptake and only secondarily on climate sensitivity. This means that for the next few decades constraints on warming are more

likely to be inferred from warming in the recent past than those on climate sensitivity itself. Further into the future, as warming increases, sensitivity becomes a controlling factor. Hence it was recommended that a vigorous program of extending and synthesizing the instrumental record should be undertaken.

Further details of the discussion and recommendations of the three breakout groups are given in the full report of the Exeter meeting by Senior et al (2004).

References

- Murphy J M. *et al.* (2004) Quantification of modelling uncertainties in a large ensemble of climate change simulations. *Nature*, to appear.
- Senior, C.A., Wielicki, B.A., McAvaney, B. and Boer, G. J., Report on the joint WCRP CFMIP/IPCC expert meeting on ‘Climate Sensitivity and Feedbacks’. Briefing material for IPCC WG1 workshop on Climate Sensitivity. See Annex 5 in this volume.
- Stainforth, D. A. *et al.* (2004). Evaluating uncertainty in the climate response to changing levels of greenhouse gases. Submitted to *Nature*.

Annex 2: Extended Abstracts

Observational Constraints on Cloud Feedback and Climate Sensitivity

Keith D. Williams

Hadley Centre, Met Office, UK

The radiative feedback from cloud remains one of the largest uncertainties in determining the climate sensitivity. The IPCC Third Assessment Report of Working Group I (Cubasch *et al.*, 2001), states that “the sign of the net cloud feedback is still a matter of uncertainty, and the various models exhibit a large spread”. Traditional methods of evaluating clouds in general circulation models (GCMs) have focused on comparing multi-annual mean maps of cloud variables (typically cloud radiative forcing or total cloud amount) with satellite observations. However, the ability to simulate the time-mean geographical distribution of present-day cloud may not offer a stringent constraint on the cloud radiative feedback under climate change. For example, a model might give a poor present-day simulation of cloud, when maps of cloud amount are compared with satellite data, because the inter-tropical convergence zone (ITCZ) is misplaced, and yet contain a physically-based cloud scheme which would provide the correct radiative feedback under climate change. Conversely, a compensation of errors might result in a model producing a realistic simulation of mean present-day cloud, but not including processes which are important for an accurate climate change simulation. A more appropriate cloud evaluation may, therefore, involve a methodology more closely related to the principal cloud processes at work.

There have been several studies which have investigated the use of compositing techniques to isolate the effect of particular processes. Bony *et al.* (1997) investigate the relationship between cloud and sea surface temperature (SST) within particular dynamical regimes, identified by 500 hPa vertical velocity over the tropical oceans; whilst Ringer and Allan (2004) composite model and observational data over the same region by vertical velocity and SST in order to evaluate different versions of a GCM. Norris and Weaver (2001) produce histograms of the amount of cloud at different heights and optical depths from the International Satellite Cloud Climatology Project (ISCCP) in three vertical velocity categories over the mid-latitude oceans. Tselioudis *et al.* (2000) also present such cloud histograms, but binned by the pressure at sea level rather than vertical velocity. Following studies by, for example, Klein and Hartmann (1993) and Slingo (1980) which identify relationships between low cloud amount and lower tropospheric stability, Weaver and Ramanathan

(1997) and Weaver (1999) composite cloud observations by vertical velocity and also by various measures of lower tropospheric stability.

These studies have generally investigated cloud processes using observational/re-analysis data and/or simulations of present-day climate from a GCM. There have been relatively few papers which have used such techniques to identify whether these relationships are important for determining the cloud response to climate change. Bony *et al.* (2004) and Williams *et al.* (2003) (hereafter W03) have proposed possible methodologies for evaluating cloud response to climate change in GCMs. The work by Bony *et al.* (2004) is described in a separate extended abstract (S. Bony: Use of observations to constrain cloud feedbacks and climate sensitivity) and hence, will not be repeated here.

W03 propose identifying key relationships between cloud changes in response to increasing greenhouse gases and changes in other atmospheric variables. If the same relationships exist in simulated spatio-temporal variability of the present day climate then the latter can be evaluated using satellite data, and the reliability of the cloud response assessed. In HadSM4 (Hadley Centre Slab Model version 4), W03 find that the high cloud response to a doubling of CO₂ over the tropical oceans is mainly associated with a change in the 500 hPa vertical velocity (more high cloud in areas of increased ascent/reduced descent and vice versa) (Fig 1a). This relationship is also present in spatio-temporal variability of present day climate, suggesting that the same fundamental processes may be responsible (Fig 1b). In this case, the variability in the present day simulation may be evaluated against observational data from the International Satellite Cloud Climatology Project (ISCCP) (Fig 1c). The model cloud amounts are from the ISCCP simulator (<http://gcssdime.gis.nasa.gov/simulator.html>), which produces special diagnostics to take account of cloud overlaps and replicate what is seen from space (Klein and Jakob, 1999; Webb *et al.*, 2001), allowing direct comparison with the observational data. The basic relationship between high cloud and vertical velocity in the model may be seen in the observations, however the cloud response in the model appears to be a little too large in those regimes with the largest changes in vertical velocity. W03 also find a relationship between the change in low cloud and the change in SST relative to the change in tropical mean

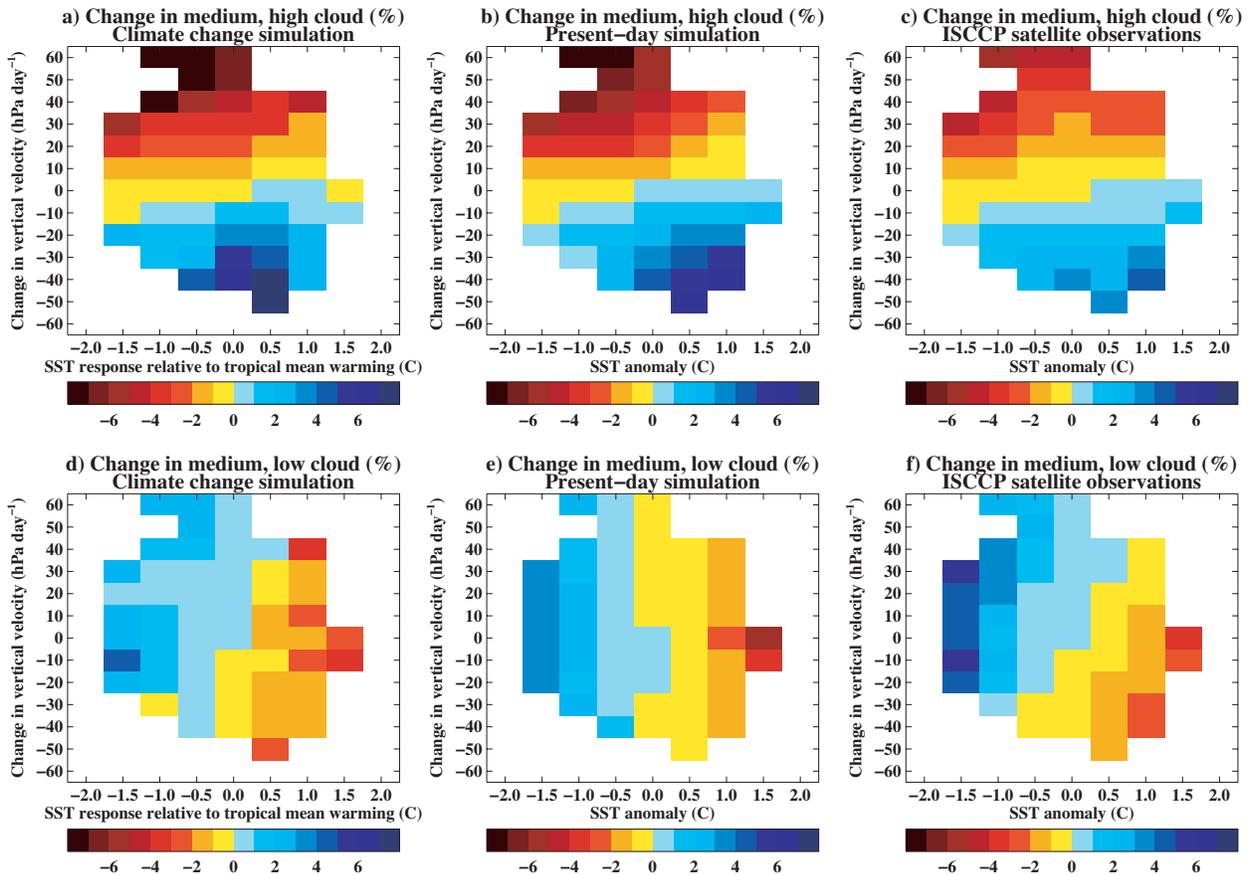


Figure 1. Change in: a-c) medium thickness high cloud amount; d-f) medium thickness low cloud amount (%). a, d) Change in cloud amount in response to doubling CO_2 in HadSM4 composited by the change in vertical velocity and change in SST relative to the tropical mean warming; b, e) Change in cloud amount in HadAM4 in response to spatio-temporal SST and vertical velocity anomalies; c, f) Change in ISCCP observed cloud amount composited by SST and vertical velocity anomalies from ECMWF re-analysis.

SST. In response to increased CO_2 , there is a reduction in cloud over those regions which have warmed more than the tropical mean and an increase in low cloud where the warming is smaller than the tropical mean. The lowcloud response to SST and vertical velocity anomalies is similar in the present day simulation and again, compares well with observations, although the data suggest that there should be stronger relationship between low top cloud and vertical velocity than simulated by the model (Fig 1d-f). It should be noted that most points lie in the bins near the origin of these plots, hence the modest cloud response in this part of the domain may contribute more to the global mean change than the more extreme response in the peripheral bins. Although the compositing shown in Fig 1 yields some success in relating cloud response to climate change over the tropical oceans to present day variability in the Hadley Centre GCM, in general, it is difficult to identify a small number of key dynamic and thermodynamic variables which together, define the climate change cloud response at a global scale.

Jakob and Tselioudis (2003) use a statistical clustering technique on ISCCP cloud histograms over the Tropical Western Pacific (TWP) to objectively identify cloud regimes. This offers the potential advantage over other compositing methodologies that explicit identification of a variable (or variables) to distinguish a regime is not required. Williams *et al.* (2004) (hereafter W04) extend the work of Jakob and Tselioudis (2003) by applying this technique to ISCCP histograms over four different geographical regions, which are known to contain different cloud regimes, and apply the same clustering to ISCCP simulator diagnostics in two versions of the the Hadley Centre GCM for present-day and climate change simulations. W04 evaluate whether the GCM can accurately reproduce the characteristics of observed cloud regimes, and whether the simulated cloud response to climate is mainly due to a change in the frequency of occurrence of regimes or a change in the cloud properties of the regimes (or a combination of the two). The work is illustrated here by examining a particular regime over

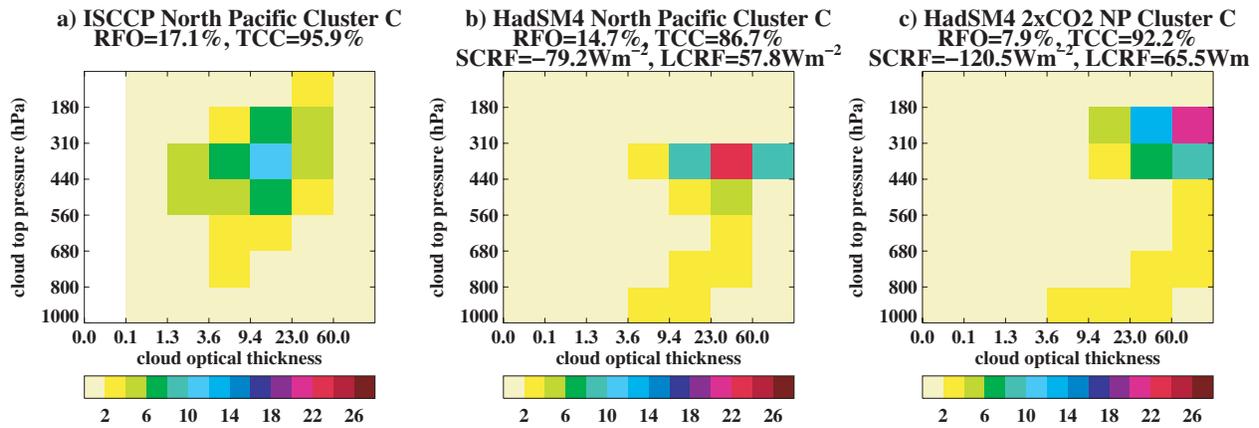


Figure 2. An example cloud top pressure - cloud optical depth cluster centroid resulting from spatio-temporal clustering 5 years of: a) daily ISCCP data; b) daily ISCCP simulator output from HadSM4 present day simulation; c) daily ISCCP simulator output from HadSM4 simulation with the CO₂ concentration doubled. In each case, the cluster shown is characteristic of the mid-latitude frontal cloud regime. Colours indicate cloud amount (%) in each cloud top pressure - optical depth category. Also shown is the relative frequency of occurrence (RFO) of the cluster, total cloud cover (TCC) of the cluster centroid and mean area weighted shortwave and longwave cloud radiative forcing (SCRF & LCRF) from the grid points included in each cluster.

the North Pacific region. Clustering of the ISCCP data over the North Pacific yields five reliable clusters, one of these, which occurs 17% of the time (the third most common), contains predominantly high topped cloud which is optically thick (Fig 2a). This regime is associated with mid-latitude frontal systems. The HadSM4 GCM simulates this cluster reasonably well, although the cloud is a little too thick and the regime is simulated too infrequently (Fig 2b). In response to climate change, the cloud in this regime becomes optically thicker and the cloud tops increase in altitude. This leads to a strengthening of the average shortwave and, to a lesser extent, longwave cloud radiative forcing components for the cluster. However, the frequency of occurrence of the cluster is halved (Fig 2c). The response of this regime is consistent with the analysis of Carnell and Senior (1998) who find, for an earlier version of the Hadley Centre model, a reduction in the number of mid-latitude storms, but an increase in their intensity, when greenhouse gases are increased. For this regime, and for most of the clusters investigated by W04 in the different regions, it is found that both changes in the frequency of occurrence and a change in the cloud properties within a cloud regime contribute to the cloud response to climate change.

W04 find that the same basic processes occur in the response of the mid-latitude frontal regime (and several other clusters) in the other version of the model they examine (HadSM3), although a differing balance of the magnitude of response from the different regimes leads to net CRF changes of opposing sign over the North Pacific in the two models. It would be very interesting to determine whether such processes occurred generally in response to increased CO₂ in different GCMs, in

which case the difference in net CRF response would simply be differences in the comparative magnitude of each process, or whether these responses are specific to the Hadley Centre model with other mechanisms determining the cloud response in other models. The Cloud Feedback Model Inter-comparison Project (CFMIP) (McAvaney and Le Treut (2003); <http://www.cfmip.net>) provides an ideal opportunity to test methodologies, such as the ISCCP clustering and other compositing techniques, in various GCMs. Through programmes such as CFMIP, which aim to evaluate several models against observations in a mechanistic framework, demonstrated to be related to climate change response, a greater understanding of simulated cloud radiative response to climate change might be gained and hopefully, uncertainty in climate sensitivity reduced.

References

- Bony, S., K.-M. Lau, and Y. C. Sud, 1997: Sea surface temperature and large scale circulation influences on tropical greenhouse effect and cloud radiative forcing. *J. Climate*, **10**, 2055–2077.
- , J.-L. Dufresne, H. Le Treut, J.-J. Morcrette, and C. A. Senior, 2004: On dynamic and thermodynamic components of cloud changes. *Clim. Dyn.*, **22**, 71–86 doi:10.1007/s00382-003-0369-6.
- Carnell, R. E., and C. A. Senior, 1998: Changes in mid-latitude variability due to increasing greenhouse gases and sulphate aerosols. *Clim. Dyn.*, **14**, 369–383.
- Cubasch, U., G. A. Meehl, G. J. Boer, R. J. Stouffer, M. Dix, A. Noda, C. A. Senior, S. C. B. Raper, and K. S. Yap, 2001: Projections of future climate change. *Climate change 2001: The scientific basis*.

- Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, Houghton, J. T., Y. Ding, D. J. Griggs, M. Noguer, P. van der Linden, X. Dai, K. Maskell, and C. I. Johnson, Eds., Cambridge University Press, 525–582.
- Jakob, C., and G. Tselioudis, 2003: Objective identification of cloud regimes in the Tropical Western Pacific. *Geophys. Res. Lett.*, **30**(21), doi:10.1029/2003GL018367.
- Klein, S. A., and D. L. Hartmann, 1993: The seasonal cycle of low stratiform clouds. *J. Climate*, **6**(8), 1587–1606.
- , and C. Jakob, 1999: Validation and sensitivities of frontal clouds simulated by the ECMWF model. *Monthly Weather Review*, **127**(10), 2514–2531.
- McAvaney, B. J., and H. Le Treut, 2003: The cloud feedback intercomparison project: (CFMIP). In CLIVAR Exchanges - supplementary contributions. **26**, March.
- Norris, J. R., and C. P. Weaver, 2001: Improved techniques for evaluating GCM cloudiness applied to the NCAR CCM3. *J. Climate*, **14**, 2540–2550.
- Ringer, M., and R. Allan, 2004: Evaluating climate model simulations of tropical cloud. *Tellus. A*, In press.
- Slingo, J. M., 1980: A cloud parametrization scheme derived from GATE data for use with a numerical model. *Q. J. R. Meteorol. Soc.*, **106**, 747–770.
- Tselioudis, G., Y. Zhang, and W. B. Rossow, 2000: Cloud and radiation variations associated with northern midlatitude low and high sea level pressure regimes. *Journal of Climate*, **13**(2), 312–327.
- Weaver, C. P., and V. Ramanathan, 1997: Relationships between large-scale vertical velocity, static stability, and cloud radiative forcing over Northern hemisphere extratropical oceans. *J. Climate*, **10**, 2871–2887.
- , 1999: The interactions among cyclone dynamics, vertical thermodynamic structure and cloud radiative forcing in the North Atlantic summertime storm track. *J. Climate*, **12**, 2625–2642.
- Webb, M., C. Senior, S. Bony, and J. J. Morcrette, 2001: Combining ERBE and ISCCP data to assess clouds in the Hadley Centre, ECMWF and LMD atmospheric climate models. *Clim. Dyn.*, **17**, 905–922.
- Williams, K. D., M. A. Ringer, and C. A. Senior, 2003: Evaluating the cloud response to climate change and current climate variability. *Clim. Dyn.*, **20**, 705–721 doi:10.1007/s00382-002-0303-3.
- , C. A. Senior, A. Slingo, and J. F. B. Mitchell, 2004: Towards evaluating cloud response to climate change using objective identification of cloud regimes. *Clim. Dyn.*, Submitted.

Annex 3:

Scoping Document for the IPCC Working Group I Workshop on Climate Sensitivity

IPCC Working Group I (WG I) will hold a workshop on the topic of Climate Sensitivity in 2004 as a major keystone in activities preparing for the WG I contribution to the IPCC Fourth Assessment Report (AR4).

Background

One of the most important parameters in climate science is the ‘climate sensitivity’, broadly defined as the global mean temperature change ($^{\circ}\text{C}$) for a given forcing, often that of a doubling of atmospheric carbon dioxide. Climate sensitivity has played a central role throughout the history of IPCC in interpretation of model outputs, in evaluation of future climate changes expected from various scenarios, and it is closely linked to attribution of currently observed climate changes. An ongoing challenge to models and to climate projections has been to better define this key parameter, and to understand the differences in computed values between various models. Throughout the last three IPCC assessments this basic parameter of the Earth’s climate system has been estimated as being in the range 1.5 to 4.5 $^{\circ}\text{C}$ (i.e., uncertain by a factor of three), making this parameter central to discussions of uncertainty in climate change.

WG I is concerned to sharpen understanding of the differences between general circulation models used in climate change research.

Currently the primary reason for the substantial range in model based estimates of climate sensitivity is widely believed to be differences in their treatment of feedbacks – particularly cloud feedbacks, but systematic intercomparisons have not been done to confirm that this is so for the current generation of models. Within international climate modeling projects, the development of new models together with both formal and informal model intercomparison exercises that are currently taking place by various groups suggest that a renewed focus on the reasons for different model estimates of climate sensitivity may be particularly useful at this time.

In addition, some recent studies suggest that new insights into the likely range of climate sensitivity may be possible through comparisons of models and observational data – both contemporary and historical or paleoclimatic. Observation/model intercomparisons will be a special focus of this workshop.

Other recent studies raise issues regarding the limitations of applicability of forcing/response relationships in the climate system - such as questions regarding the degree of predictability of climate and its relevance for estimates of climate sensitivity, and the degree to which forcings such as those due to solar, well-mixed greenhouse gases, or aerosols may produce different responses. A review of these questions about the interpretation of climate sensitivity could also sharpen scientific understanding and would hence be of benefit to the WG I AR4.

In summary, there is broad interest for a carefully planned workshop on climate sensitivity. Given the importance of the climate sensitivity parameter, it is likely that the outcome of this workshop will provide a major focus for the discussion and treatment of climate models in the WG I contribution to AR4.

Aims

The aims of the climate sensitivity workshop would be to:

- Evaluate a range of climate model results so as to relate different climate sensitivity estimates to differences descriptions of physical processes, particularly those related to atmospheric water vapor, clouds, lapse rate changes, ocean heat uptake, treatment of evapotranspiration, land-atmosphere coupling, etc.;
- Obtain a more comprehensive picture of the relationships between climate sensitivity and other model features such as resolution, numerical approach, radiative transfer parameters, etc.;
- Consider how current, historical, and/or paleoclimatic data can aid in the determination of the likely range of climate sensitivity;
- Improve the understanding of the interpretation and limits of the climate sensitivity concept, including for example possible dependencies upon different forcing agents, predictability questions, and transient and steady-state responses;
- Start a process towards objective assessment to critically determine whether the range 1.5 to 4.5°C remains appropriate in the AR4 – e.g. by defining criteria that may assist in the evaluation of results from many different climate models.

Approach and Timetable

Given the range of issues to be considered and the commitment that would be required from major modeling groups around the world, the process will be structured by a broad-based scientific steering group. Planning for the workshop will be carried out by the steering group. Organizational support for the meetings, and production of a workshop report will be carried out by the WG I TSU.

In order to include a carefully constructed intercomparison of climate model results as part of the proposed workshop, a preliminary expert meeting will foster explicit analysis of feedbacks, in collaboration with WCRP activities on this topic. The expert meeting is currently planned for April 2004 in Exeter, UK.

The workshop will take place on July 26-29, 2004 in Paris, France.

Annex 4:

List of Participants

Patricio Aceituno

Department of Geophysics
University of Chile
CHILE

Epiphane Dotou Ahlonsou

Service Météorologique National
REPUBLIC OF BENIN

Filipe Aires

Laboratoire de Météorologie Dynamique du CNRS
FRANCE

Abdelkader Allali

Ministry of Agriculture, Rural Development & Forestry
MOROCCO

Richard Allan

Environmental Systems Science Centre (ESSC)
University of Reading
UNITED KINGDOM

Myles Allen

Atmospheric, Oceanic and Planetary Physics (AOPP)
University of Oxford
UNITED KINGDOM

Caspar Ammann

National Center for Atmospheric Research (NCAR)
USA

Natasha Andronova

Department of Atmospheric Sciences
University of Illinois @ Urbana-Champaign
USA

George Boer

Canadian Centre for Climate Modelling and Analysis
Meteorological Service of Canada
University of Victoria
CANADA

Sandrine Bony

Laboratoire de Météorologie Dynamique du CNRS
FRANCE

Pascale Braconnot

Laboratoire des Sciences du Climat et de
l'Environnement
L'Orme des Merisiers
FRANCE

Anthony Broccoli

Department of Environmental Sciences
Rutgers University
USA

Jean-Yves Caneill

EDF R&D
TESE Department
FRANCE

Robert Cess

Institute for Terrestrial and Planetary Atmospheres
Stony Brook University
USA

John Charlery

Department of Computer Science
University of the West Indies
BARBADOS

Jens Hesselbjerg Christensen

Danish Meteorological Institute
DENMARK

Matthew Collins

Hadley Centre for Climate Prediction and Research
Met Office
UNITED KINGDOM

William Collins

National Center for Atmospheric Research (NCAR)
USA

Robert Colman

Bureau of Meteorology Research Centre
AUSTRALIA

Thomas Crowley

Department of Earth and Ocean Sciences
Duke University
USA

Hervé Douville

Meteo France CNRM
FRANCE

Jean-Louis Dufresne

Laboratoire de Météorologie Dynamique du CNRS
FRANCE

Seita Emori

Integrated Modeling Research Program
Frontier Research System for Global Change
JAPAN

David Fisher

Geological Survey of Canada
CANADA

Donald Forbes

Geological Survey of Canada
CANADA

Piers Forster

Department of Meteorology
University of Reading
UNITED KINGDOM

Raquel Francisco

PAGASA, Weather Bureau
PHILIPPINES

Qiang Fu

Atmospheric Sciences
University of Washington
USA

Carlos Gay García

Centro de Ciencias de la Atmosfera
Universidad Nacional Autonoma de Mexico (UNAM)
MEXICO

Michael Ghil

Department of Atmospheric Sciences and Institute of
Geophysics and Planetary Physics
University of California, Los Angeles
USA

Marc Gillet

Observatoire National sur les Effets du Réchauffement
Climatique
FRANCE

Marco Giorgetta

Max Planck Institute for Meteorology
GERMANY

Jonathan Gregory

Department of Meteorology
University of Reading
UNITED KINGDOM

Alex Hall

Department of Atmospheric Sciences
University of California, Los Angeles
USA

Stephane Hallegatte

International Center for Research on Environment and
Development (CIRED)
FRANCE

Gabriele Hegerl

Division of Earth and Ocean Sciences
Duke University
USA

Isaac Held

NOAA Geophysical Fluid Dynamics Laboratory
(GFDL)
USA

Trond Iversen

Department of Geophysics
University of Oslo (UiO)
NORWAY

Eystein Jansen

University of Bergen
Department of Earth Sciences and Bjerknnes Centre for
Climate Research
NORWAY

Fortunat Joos

Physikalisches Institut - Klima- und Umweltphysik
University of Bern
SWITZERLAND

Sylvie Joussaume

National Institute for Sciences of the Universe (INSU)
FRANCE

Jean Jouzel

Institut Pierre Simon Laplace
FRANCE

Eigil Kaas

Climate Research Division
Danish Meteorological Institute
DENMARK

Lucka Kajfez-Bogataj

University of Ljubljana
Biotechnical Faculty
SLOVENIA

Vladimir Kattsov

Voeikov Main Geophysical Observatory
RUSSIA

Richard Kerr

Science Magazine
USA

Tahl Kestin

IPCC WGI TSU
USA

Jeff Kiehl

National Center for Atmospheric Research (NCAR)
USA

Masahide Kimoto

Center for Climate System Research University of
Tokyo
JAPAN

Stefan Kinne

Max Planck Institute for Meteorology
GERMANY

Akio Kitoh

Meteorological Research Institute
Japan Meteorological Agency
JAPAN

Reto Knutti

Physics Institute
University of Bern
SWITZERLAND

Won-Tae Kwon

Meteorological Research Institute (METRI)
Korean Meteorological Administration
REPUBLIC OF KOREA

Hervé Le Treut

Laboratoire de Météorologie Dynamique du CNRS
FRANCE

Laurent Li

Laboratoire de Météorologie Dynamique du CNRS
FRANCE

Scott Longmore

IPCC WGI TSU
USA

Armella LONGREZ

Institut Pierre Simon Laplace
FRANCE

Syukuro Manabe

Program in Atmospheric and Oceanic Sciences
Princeton University
USA

Martin Manning

IPCC WGI TSU
USA

Valérie Masson-Delmotte

Laboratoire des Sciences du Climat et de
l'Environnement
FRANCE

Bryant McAvaney

Bureau of Meteorology Research Centre
AUSTRALIA

Mack McFarland

DuPont Fluoroproducts
USA

Gerald Meehl

National Center for Atmospheric Research (NCAR)
USA

Valentin Meleshko

Voeikov Main Geophysical Observatory
RUSSIA

Ronald Miller

NASA Goddard Institute for Space Studies
USA

Emmanuel Mpeta

Tanzania Meteorological Agency
TANZANIA

James Murphy

Hadley Centre for Climate Prediction & Research
Met Office
UNITED KINGDOM

Akira Noda

Meteorological Research Institute
Japan Meteorological Agency
JAPAN

Toru Nozawa

National Institute for Environmental Studies (NIES)
JAPAN

Bartolomé Orfila-Estrada

Instituto Nacional de Meteorología (INM)
SPAIN

Bette Otto-Bliesner

National Center for Atmospheric Research (NCAR)
USA

Jonathan Overpeck

Institute for the Study of Planet Earth
University of Arizona
USA

Beenay Pathack

Meteorological Services
MAURITIUS

Michel Petit

FRANCE

Serge Planton

Météo-France CNRM
FRANCE

Stefan Rahmstorf

Potsdam Institute for Climate Impact Research (PIK)
GERMANY

Jouni Räisänen

Department of Physical Sciences
University of Helsinki
FINLAND

V. Ramaswamy

NOAA Geophysical Fluid Dynamics Laboratory
(GFDL)
USA

David Randall

Department of Atmospheric Science
Colorado State University
USA

Sarah Raper

Alfred Wegener Institute
GERMANY

Alan Robock

Department of Environmental Sciences
Rutgers University
USA

Rémy Roca

Laboratoire de Météorologie Dynamique du CNRS
FRANCE

Jean-François Royer

Météo France CNRM
FRANCE

Robert Sausen

Institute of Atmospheric Physics
Deutsches Zentrum für Luft- und Raumfahrt e.V. (DLR)
GERMANY

Michael Schulz

Laboratoire des Sciences du Climat et de
l'Environnement
FRANCE

Cath Senior

Hadley Centre for Climate Prediction and Research
Met Office
UNITED KINGDOM

Madan Shrestha

Department of Hydrology & Meteorology
NEPAL

Silvina Alicia Solman

National Council of Research and Development of
Argentina -CIMA (CONICET-UBA)
ARGENTINA

Susan Solomon

Co-Chair, IPCC WGI
NOAA Aeronomy Laboratory
USA

Richard Somerville

Scripps Institution of Oceanography
University of California, San Diego
USA

Asgeir Sorteberg

Bjerknes Centre for Climate Research
NORWAY

J. Srinivasan

Centre for Atmospheric and Oceanic Sciences and
Indian Institute of Science
INDIA

David Stainforth

Atmospheric, Oceanic and Planetary Physics (AOPP)
University of Oxford
UNITED KINGDOM

Thomas Stocker

Physikalisches Institut - Klima- und Umweltphysik
Universität Bern
SWITZERLAND

Annex 4: List of Participants

Ron Stouffer

NOAA Geophysical Fluid Dynamics Laboratory
(GFDL)
USA

Olivier Talagrand

Laboratoire de Météorologie Dynamique du CNRS
France

Melinda Tignor

IPCC WGI TSU
USA

George Tselioudis

NASA Goddard Institute for Space Studies and
Columbia University Department of Applied Physics
USA

Rob van Dorland

Royal Netherlands Meteorological Institute (KNMI)
THE NETHERLANDS

Keith Williams

Met Office
UNITED KINGDOM

Annex 5:

Report on the Joint WCRP CFMIP/IPCC Expert Meeting on ‘Climate Sensitivity and Feedbacks’

Catherine Senior¹, Bruce Wielicki², Bryant McAvaney³ and George Boer⁴

¹ *Hadley Centre, Met Office, U.K.*

² *NASA Langley Research Centre, Virginia, USA*

³ *Laboratoire de Meteorologie Dynamique du CNRS, Paris, France*

⁴ *Canadian Centre for Climate Modelling and Analysis, University of Victoria, Canada*

Background

An expert meeting on ‘Climate Sensitivity and feedbacks’ was held at the Met Office in Exeter, U.K from April 19th-22nd, 2004. The meeting was jointly arranged by IPCC WG1 and the WCRP CFMIP (Cloud Feedback Inter-comparison project: <http://www.cfmip.net>), and acted both as a first meeting of CFMIP and as an expert meeting for IPCC WG1 to develop scientific questions relevant to the intergovernmental workshop on Climate Sensitivity, July 26th-29th in Paris. The Exeter meeting had as its main focus the role of clouds and cloud feedbacks in determining climate sensitivity. The meeting had 3 major sessions on ‘Feedback Methods’, ‘Cloud: Validation and feedbacks’ and ‘CFMIP results and Plans’. In addition a number of key speakers gave presentations at a Royal Meteorological Society meeting on ‘Climate Sensitivity and Feedbacks’ held during the meeting. On the final day of the meeting, three breakout groups were asked to discuss topics that had arisen during the meeting. These topics were chosen to be issues on which we might be able to make recommendations to IPCC WG1 as ways forward to addressing some of the key scientific questions raised in the IPCC WG1 Workshop on Climate Sensitivity ‘Scope and Aims’ document. Below we summarise the topics of the breakout groups and the recommendations put forward.

Summary of Breakout Group (BOG) discussions

1 *BOG1: Climate Feedback Metrics*

During the meeting, a number of speakers had talked about the possibility of constraining climate sensitivity using a basket of measures (or metrics) aimed at validating model representations of present day climate. BOG1 started by reviewing the utility of current experimental designs (notably from the CFMIP project) for such a purpose and looked at what data would be needed to allow the necessary metric tests. The focus was particularly on clouds, although it was recognized that such an approach could be broadened as required. A number of climate model ensembles (the so-called QUMP ensemble at the Hadley Centre (Murphy et al., 2004), the climateprediction.net ensemble (Stainforth et al, 2004) and the forthcoming CFMIP ensemble are available (or will be soon) and BOG1 propose a way forward for using such ensembles that aims to relate errors in model simulations (when compared to observations) to uncertainty in climate prediction. The ultimate goal being to reduce uncertainty in climate sensitivity through application of such carefully constructed metrics to a range of models.

2 *BOG2: CFMIP*

Those involved in CFMIP discussed the plans for the project with particular reference to IPCC WG1, and highlighted a number of key areas. A recommendation was made that the CFMIP data be used for research into developing climate metrics, notably related to clouds (stratocumulus and the link between upper tropospheric humidity and cirrus cloud were considered most urgent). Also it was recommended that the ISCCP-simulator (a software package that can be applied to model output in order to produce quantities that can be directly compared to quantities retrieved from the International Satellite Cloud Climatology Project) be included in IPCC AR4, AMIP and CMIP runs. It will be recommended to groups involved in CFMIP to diagnose instantaneous radiative forcing (as will be available for AR4 experiments) and to use partial radiative perturbations for analyzing cloud feedbacks in addition to the more common approach of looking at changes in cloud radiative forcing. A novel area of research for CFMIP is developing methodologies (e.g. compositing and clustering) to evaluate aspects of clouds in models that can be shown to be important for cloud response to climate change. It is recommended that such techniques should be applied to a wider range of models, e.g. those in AMIP, CMIP and AR4. This could be best achieved if modeling groups who submit data to AR4 can be encouraged to also participate in CFMIP.

3 *BOG3: Climate sensitivity and response*

The third BOG discussed the relative roles and importance of transient and equilibrium climate sensitivity particularly to policy makers. Recommendations were that modeling groups should provide the temporal evolution of the effective (or transient) climate sensitivity to IPCC AR4 and should be encouraged to analyse the spatial and temporal structure of the feedbacks under a range of IPCC forcings. There was discussion of the relationship between climate sensitivity, ocean heat uptake and temperature response and ways of constraining climate sensitivity and temperature response using the past behavior of the climate system. One approach is the use of climate perturbations that are comparatively well known. A recommendation was made that IPCC AR4 coupled models include the simulation of the climate perturbation associated with the Mount Pinatubo eruption, paying particular attention to the nature and timescale of the simulated response, the role of ocean heat uptake and the bounds that can be placed upon climate sensitivity. Another approach is the use of observed trends (both in the instrumental record and in palaeoclimate reconstructions) to constrain climate sensitivity. A recommendation was made that studies of past forced climate variations should be pursued but will require carefully developed datasets of forcing and response. Finally constraining temperature evolution was discussed, as for small changes this is primarily dependent on forcing and ocean heat uptake and only secondarily on climate sensitivity. This means that for the next few decades constraints on warming are more likely to be inferred from warming in the recent past than those on climate sensitivity itself. Further into the future, as warming increases, sensitivity becomes a controlling factor. Hence it was recommended that a vigorous program of extending and synthesizing the instrumental record should be undertaken.

Further details of the discussion and recommendations of the three breakout groups are given in the attached reports for each breakout group

BOG 1: Climate Feedback Metrics

Recommendations

1 Overall Modelling Simulation Strategy

The BOG 1 reviewed CFMIP experimental design with the goal of identifying ways of leveraging this experimental activity to address our charge. Questions that were discussed but not fully addressed included whether these experiments should be equilibrium-like experiments or should include the observed natural and anthropogenic forcing record (i.e., a transient slab experiment). Group discussion favoured the equilibrium strategy which is also more consistent with the CFMIP experimental design. We noted, however, that an attempt should be made to allow some of the CFMIP runs to better simulate an ENSO forcing by imposing spatial Q-flux patterns.

We concluded that the same cloud/climate metrics would be needed from the range of studies proposed for AMIP (C20C using observed SST and forcing), CFMIP, as well as the fully coupled IPCC runs. We recommend that the same set of data (at a minimum) be collected as planned for the CFMIP, with the selected additions shown below in item 2. There is a concern that the ISCCP cloud pressure/optical depth statistics might not be included in the IPCC runs and we strongly recommend this data be included for study.

We noted that each of the different simulation methods brings different strengths and weaknesses in evaluating climate simulations versus observations. AMIP simulations provide more accurate boundary conditions such as ENSO, but break the surface energy budget feedbacks. CFMIP slab runs bring in the surface flux feedbacks, but are weak for ENSO events and are missing most ocean physics. Fully coupled IPCC 20th century runs enable the full ocean/atmosphere interaction but have limited numbers of simulations. We also considered the QUMP (Murphy et al, 2004) and climateprediction.net (Stainforth et al., 2004) efforts for large ensembles of Perturbed Physics Ensembles (PPEs). These are considered a key element of mapping errors in model simulations to uncertainty in climate sensitivity. These are dealt with in item 3 below.

In summary, tests of climate models versus observations will be required in a wide range of time/space scales and methods to handle the complete range of climate feedbacks. Ensembles of model runs will be key to determining natural variability, since all comparisons to observations are to a single realization of the Earth's climate system. Significance of climate metrics for model minus observations must have both observational error statistical distributions as well as model climate noise statistical distributions. Both must be defined at the relevant range of time/space scales to handle a complete range of climate feedbacks. While many "necessary" climate model tests are known, there currently exists no set of "sufficient" climate model tests. This remains a critical and active area of research, and this report is a further step toward a complete set of model tests.

2 Model Run Data Needed for Metrics: AMIP, CFMIP, IPCC

We reviewed the CFMIP planned variables and time/space scales being saved and considered whether these were adequate for cloud feedback studies in general. While the general specifications looked very good, we had a few key additions:

- a) Aerosol information if the model uses aerosols in the run. Properties should include visible optical depth, aerosol layer mean height, aerosol type/species, particle size and assumed single scatter albedo. If type/species specify all optical properties but optical depth, then these can be specified once for each type/species
- b) 3-hourly sampling for the diurnal average variables is needed to study diurnal cycles of cloud. It would be sufficient to provide 3 hourly variables for the seasonal months of January, April, July, and October in 4 sample years. The selection of months and years is done to minimize data volume while retaining sufficient samples of diurnal cycle and inter-annual variability.
- c) Note that ISCCP cloud pressure/optical depth cloud statistics are required at all time/space scales included in the planned CFMIP data

3 Relating Model Metrics to Climate Sensitivity

One of the major shortcomings of all climate prediction assessments has been the inability to objectively and unambiguously relate the errors in model simulations (as judged by comparisons to observations) to uncertainty in climate sensitivity, i.e.

Climate Sensitivity Uncertainty = Function (Model Error versus Observations)

One of the problems immediately obvious (and why this has not been accomplished to date) is that climate model error metrics vary with a) model variable b) spatial scale c) time scale d) single variable errors versus multiple correlated variable errors. The fundamental nonlinear nature of the climate system guarantees that this function is unlikely to be simple. To date attempts have been ad-hoc and have varied widely depending on physical feedback mechanism, variable, time and space scale. What is needed is a more general approach.

QUMP and climateprediction.net offer a new opportunity to attack this outstanding problem in climate assessment and climate science. A strategy was discussed by the working group and will be explored in the coming year. A brief description and outline of steps is given below.

The methodology is based on the following concept. Consider any two climate models with different physics (could be parameterization, resolution, etc). Each model is run for control climate (fixed boundary conditions) and for perturbed boundary conditions. The example we will use here is the slab ocean runs for fixed CO₂ and for doubled CO₂. Consider each model as a real planet with perfectly known physics (the model formulation) and perfectly known observations (model output fields). The key is to consider planet 1 as a model of planet 2. We in essence use

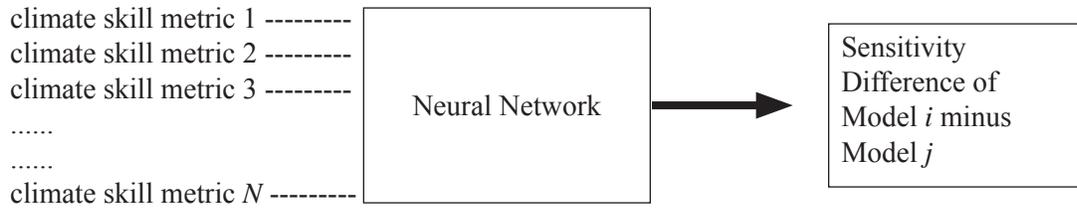
the climate sensitivity of model 1 to be a predictor of climate sensitivity for model 2. Since the models have different physics, the prediction will have error. But we can also compare the perfect observations of planet 2 (model 2 output) to the prediction of model 1. There are also errors in these fields. Analogy here is that we want to learn how to use differences in model 1 minus model 2 output fields (climate metrics) to predict the sensitivity difference of model 1 minus model 2. Model 2 is the real earth, Model 1 is the prediction model of Model 2. The next key is to vary all of the uncertain physical parameterizations in the climate system within reasonable ranges. This might be 50 tuneable parameters, each varied both one at a time and in combination, so that 100s to 1000s of different physical models are generated. The tuneable parameters are chosen to be broad and to encompass most if not all of the known climate issues: surface physics, ocean physics, atmosphere physics, ice sheet physics, sea ice physics, chemistry, etc. We now have 1000s of planets, each somewhat like Earth, but each with different climate feedbacks. If we run each of these models in a fixed CO₂ and doubled CO₂ version, we also have the CO₂ doubling sensitivity of each planet (model). We call this a Physics Parameter Ensemble (PPE). When selected as individual pairs each model pair is used to get one realization of a real planet and a model trying to predict the sensitivity of that planet. Note that the climate sensitivity we are trying to predict does not have to be restricted to global average temperature. We could use any climate metric as the desired sensitivity output: Europe summer precipitation, ENSO magnitude, storm track shifts, etc. We will use the example here of doubling of CO₂ temperature climate sensitivity.

3.1 Step 1: Run the PPE ensemble in fixed CO₂ and doubled CO₂

These runs must vary a wide range of physical parameterizations. In initial tests, the Hadley Centre has run 58 PPE samples and has begun studying the application of a broad Climate Prediction Skill (CPS) of about 30 parameters. They have saved a fairly complete set of climate diagnostics from this run. Climateprediction.net has run a much larger set of PPE experiments (1000s) but has much more limited climate diagnostics. They are now working to improve this set in future simulations submitted. Ultimately the types of metrics described in item 2 of this report will be desired. Because of the large number of model runs involved, a subset of these will be necessary initially.

3.2 Step 2: Select a set of climate skill metrics and train a neural net to relate Climate Prediction Skill to climate sensitivity difference

Select each possible pair of climate model runs. Use all of the climate prediction skill metrics (order 30 to begin with) for each model pair as input to the neural net. Output is the climate sensitivity difference between the two models. As discussed earlier, this mimics normal climate prediction, but where model 2 becomes the “Earth” and model 1 is the model trying to predict the “Earth’s” (model 2) sensitivity. The difference in sensitivity between the two models is the error in its prediction of sensitivity. The climate metrics differences between the two models are the error in the “observations of Earth” (model 2 output fields) versus the prediction (model 1 output fields). Do this for all model pairs. There will be 1000s to millions of such pairs depending on the total number of models M in the PPE set.

Model i predicts Model j 

Once the neural net is complete, test its accuracy in prediction of sensitivity difference between each model pair. If this accuracy is sufficient then the set of climate metrics is deemed useful, note the uncertainty of the prediction (PDF of error in sensitivity difference prediction) and move to step 3. If the prediction is not useful (say 100% 1sigma uncertainty in sensitivity difference between model i versus model j) then the set of climate metrics was insufficient. Note that in using the climate metrics, a neural net will automatically give heavier weight to more important metrics, lower weight to less important metrics, and will allow for correlations between metrics. These are all key characteristics of moving beyond a simple linear sum of errors in all parameters and time/space scales. In all cases, the key to a robust neural net prediction is a large ensemble of tests, a complete set of climate sensitivity variations, and a good set of climate metrics.

3.3 Step 3: Test Neural Network against an independent set of models: CFMIP slab runs

Since all of the PPE runs were built on parameterization changes in the UKMO climate model, it is important to verify the neural net performance against an independent set of climate models with very different approaches to parameterization. The CFMIP slab runs of fixed CO₂ and doubled CO₂ provide this independent test. Test all pairs of CFMIP model runs using the neural net developed in step 2, to predict their sensitivity differences. The accuracy of this prediction should be compared against the training set accuracy in step 2. If accuracy is similar and still within goals, proceed to step 4. If not, will need to run a large ensemble of PPE simulations using varying climate models, not just the UKMO model. Then repeat step 2.

An additional useful test of the rigor of the neural net climate sensitivity predictions is to perform a set of solar constant change experiments with the same CFMIP slab climate models. The neural net can be used to predict the model climate sensitivities in a manner similar to the primary CO₂ doubling experiment.

3.4 Step 4: Use fully coupled 20th Century IPCC runs with and without anthropogenic forcing

Use model pairs as in step 3 to test prediction capability of neural net for changing climate sensitivity between each pair of models. If this is not successful, may need to run a PPE set of fully coupled models to add in deep ocean physics. If it is successful, then move to step 5.

3.5 Step 5: Evaluate Uncertainty in IPCC model sensitivity

Determine the climate metrics comparing the IPCC 20th century runs and actual Earth observations. Note that this is the first actual use of observations in this evaluation process. All

other training and testing of the neural net has used model simulations as alternative Earth/model pairs. Their purpose was to demonstrate the accuracy and rigor of the neural net system to relate climate metric differences to climate sensitivity differences. We now use actual observations of the Earth versus one of the climate models to set the climate metrics that are input to the neural net. The neural net then converts differences between model and observed Earth to predict the difference in climate sensitivity between model and Earth. An additional test is to compare the predicted climate sensitivity of the coupled models to the actual values. Do these steps for all IPCC model runs. For each model you will have its climate sensitivity as well as a prediction of the difference between the model and the true Earth sensitivity. For K model IPCC runs, there will be K predicted sensitivity and differences. There is also a known “noise” from step 4 in the ability to use the climate metrics to predict sensitivity differences. This allows a statistical test of what level all K models are consistent with a range of the true Earth climate sensitivity. We may find that a narrow range of Earth climate sensitivities would be consistent with all of the model predictions: this then would be the most likely range of sensitivity and could be stated as a function of different percentiles of likelihood. If step 4 shows good predictability of the climate sensitivity uncertainty, but step 5 shows an inconsistent range of sensitivity plus uncertainty from the models, this result would point toward the likelihood of some unrealized physical process in the real Earth system that was not included in any of the PPE simulation variations of climate feedback processes.

3.6 Effect of Observational Noise and Climate System Internal Natural Variability

The neural network development intrinsically includes internal climate noise, at least as far as it is reasonably simulated by the climate models. The analysis proposed here can also be used, however, to test the effect of observation error on the neural net prediction of climate uncertainty. A brute force method is to Monte Carlo added noise onto the climate skill metrics, but there are more elegant ways to do this within the framework of a neural network approach.

3.7 Neural Networks and Other Analysis Tools

Neural networks are only one type of statistical tool that could be applied to evaluate the relationships discussed in sections 3.1 to 3.6. The results are not guaranteed to be unique, especially if the set of training cases is too small or if the set covers only a small subset of the range of application (lack of completeness). The suggested tests in 3.3 and 3.4 are designed to highlight such difficulties if they occur, but other statistical methods should also be evaluated, and in the end might prove more effective. Neural networks are used in the discussion here as a suggested initial approach with broad capabilities.

There is also concern that a neural network approach is a “black box” with reduced insight into the key physical relationships. This criticism suggests that alternative linear and nonlinear approximations to the full climate system must be examined in parallel with statistical approaches. But the criticism can also be reduced by selection of climate variable metrics in 3.2 based on known physical forcing and feedback mechanisms in the climate system. The major challenge for any successful method is obtaining relationships that are valid in the fully coupled nonlinear climate system. The strongest evidence and physical insight will come from

convergence of results from multiple approaches. The neural network approach described here has the potential to significantly advance this critical effort.

3.8 *Can the Earth's True Climate Sensitivity be Outside the Uncertainty Range Determined in Section 3.5?*

Yes, but it is less likely than current uncertainty estimates. To date, our climate sensitivity uncertainty estimates have relied primarily on the range in sensitivity of the current range of climate models. But the uncertainties in sensitivity predicted in section 3.5 are based in addition on the differences between the model runs and our actual observations: i.e. the bigger the differences the bigger the uncertainty. This is a new constraint on climate sensitivity that has not previously been available. For example, consider what would happen in the situation where all models are missing some fundamental cloud feedback mechanism that a) occurs in nature and b) is a large part of total cloud feedback. In this case, the true climate sensitivity might lie outside the range of any of the current climate models. The process in sections 3.1 to 3.5 adds an additional requirement, that the missing feedback c) doesn't affect significantly the observational metrics we used to test the models. Item c) is the new constraint we add in the current process in section 3.5. Past uncertainty estimates have not been able to explicitly incorporate model minus observation metrics. If the climate metrics used in 3.1 through 3.5 are physically based observational metrics for each feedback (in this case the relevant cloud properties and radiative fluxes) then a missing feedback should show up as large observation/model error, and as a result in a larger climate sensitivity uncertainty. While still not perfect, a climate feedback sufficient to cause the true climate sensitivity to be outside the range of the uncertainty estimate would now have to pass all three tests a) through c).

BOG 2: CFMIP (Cloud Feedback Model Intercomparison Project)

Recommendations

1 Metrics

An incomplete list of priority metrics on the basis of currently known modelled responses: a) a metric to determine stratocumulus (and other types of boundary layer cloud) response is most urgent. b) metrics that can help investigate the link between upper tropospheric humidity and cirrus

Recommend: Use of output of CFMIP by others as research into metrics methodologies continues.

Recommend: ISCCP simulator be installed in IPCC AR4 models (and CMIP and AMIP).

2 Radiative Forcing

CFMIP will request modelling groups to diagnose radiative forcing (from 2xCO₂ only), using the instantaneous change double radiation call approach. The forcing should be diagnosed at TOA,

surface and 200 mb, and split into SW and LW, up and down, clear-sky and all sky net fluxes. At a bare minimum at-least-5-year means of the geographical distribution should be provided. This should be linked to the WCRP (GEWEX Radiation Panel) Intercomparison of Radiation codes in Climate Models (ICRCCM).

Participants are also encouraged also to calculate the surface, TOA and 200 mb “Hansen”/ “relaxed” forcing (double CO₂ but keep SST constant; Hansen et al,1997).

Recommend: Instantaneous radiative forcing required from all participating CFMIP models (slab ocean experiment)

Recommend: Connection with WCRP ICRCCM radiation intercomparison programme.

3 Radiative Changes

Partial radiative perturbation (PRP) analysis indicates more cleanly what causes what radiative changes, but cloud radiative forcing (CRF) change is more related to observations. CFMIP encourages participants to use both, everyone is urged to keep in mind the differences between them. CFMIP will recommend standards for the detail (e.g. RH as well as q, double substitution, zonal means / maps). The use of simple models to interpret changes in CRF in the absence of PRP diagnostics will be tested using the models with PRP diagnostics. Alternative definitions of CRF which do not suffer from the cloud masking effect in the clear-sky fluxes will be sought.

Recommend: Individual modelling groups to perform PRP analysis “in house” using standardised techniques.

4 Compositing and Clustering

A novel area of research for CFMIP is developing methodologies (e.g. compositing and clustering by cloud types) to evaluate aspects of clouds in models that can be demonstrated to be important for the cloud response to climate change and hence, climate sensitivity. It is recognised that there is a need to explore links between different parts of the hydrological cycle – CFMIP should aim to develop techniques in this area e.g. storm/precipitation compositing (George Tselioudis, personal communication)

Recommend: Compositing techniques should be applied to model results submitted to AMIP and CMIP and IPCC AR4.

Recommend: Develop compositing techniques that represent the variations of combinations of cloud types that are associated with distinct meteorological states.

Recommend: AR4 participants should submit results to CFMIP.

5 Other Issues

Recommend: Reinforce that the atmospheric model used for CFMIP should be the same as that used for AMIP and CMIP.

Recommend: CFMIP should provide feedback to the GEWEX Cloud System Study (GCSS) on systematic problems in areas important for cloud response, and should also advise on experimental design for climate change studies using Cloud Resolving Models.

Advantage of idealised $\pm 2\text{K}$ runs (as well as backward compatibility) is being able to compare models' response with same warming, but it is an unrealistic one. Imposing structured SST changes (e.g. using the CMIP mean or median) would maintain that advantage while reducing the unrealism and so is recommended for the future. It could also be used for ensembles of perturbed physics, as discussed in BOG1.

Recommend: From CMIP create ensemble mean structured SST mean change and test as an alternative method to slab ocean models for evaluation cloud feedbacks.

BOG3: Climate sensitivity and response

Recommendations

1 Effective and equilibrium climate sensitivity

Climate sensitivity is generally defined as the global mean (indicated by angular brackets) surface temperature change $\langle T'_{2x} \rangle$ experienced by the climate system, or a model of the system, after it has attained a new equilibrium after a doubling of atmospheric CO_2 concentration (IPCC2001). The CO_2 perturbation to the radiation stream is measured by the *radiative forcing* $\langle f_{2x} \rangle$ and the *equilibrium sensitivity parameter* s_e links the forcing to the temperature response as

$$\langle T'_{2x} \rangle = s_e \langle f_{2x} \rangle \quad (1)$$

To the extent that the sensitivity parameter is approximately constant and independent of the nature and pattern of the radiative forcing, (1) provides a way of estimating the global mean temperature response from the magnitude of the forcing and thus broadly characterizes the system. The global climate sensitivity is also used for scaling results, for calibrating simpler models, for estimating the temperature effects of different radiative forcings, and for comparing the behaviour of different models. However, it is worth noting that it cannot be directly observed or measured.

While the climate sensitivity of a particular model is comparatively robust, values of climate sensitivity differ by as much as a factor of two among models (e.g. IPCC2001, Fig 9.18) implying that projections of global warming can also differ by that factor for the global mean and by larger amounts locally. It is important to understand why this is the case, to probe the

robustness of the sensitivity over a range of forcings, and to understand the connection between temperature response and sensitivity as the system evolves in response to radiative forcing.

The *equilibrium* climate sensitivity in (1) may be diagnostically linked in a model to a more general *effective* climate sensitivity via the energy budget (Murphy, 1995). In a more general approach, the forcing $f(\lambda, \phi, t)$ and the temperature response $T'(\lambda, \phi, t)$ are functions of location and time and are linked through the vertically integrated energy budget equation (Boer and Yu, 2003a), as

$$dh'/dt = A' + R' = A' + g + f = A + \Lambda T' + f \quad (2)$$

where $X' = X - X_0$ is the difference from the control or current climate, dh'/dt the heat storage (mainly in the ocean), A' the change in the convergence of horizontal heat transport and $R' = g + f$ the radiative perturbation which is decomposed into the radiative forcing f and the radiative feedback $g = \Lambda T'$ expressed also as a function of local temperature response and a feedback parameter $\Lambda(\lambda, \phi, t)$. The connection with (1) follows by averaging as $\langle dh'/dt \rangle = \hat{\Lambda} \langle T' \rangle + \langle f \rangle$, with $\hat{\Lambda} = \langle \Lambda T' \rangle / \langle T' \rangle$, or alternatively as

$$\langle T' \rangle = -\langle f - dh'/dt \rangle / \hat{\Lambda} = s \langle f - dh'/dt \rangle \rightarrow s_e \langle f \rangle \quad (3)$$

$$s = \langle T' \rangle / \langle f - dh'/dt \rangle \rightarrow s_e \quad (4)$$

where the arrow indicates the limit as equilibrium is approached. All terms, including the effective sensitivity parameter $s(t) = 1 / \hat{\Lambda}(t)$ are nominally functions of time. When the system reaches a new equilibrium, $\langle dh'/dt \rangle \rightarrow 0$, the effective sensitivity s becomes the equilibrium sensitivity s_e . The evolution of effective sensitivity is seen in some (Senior and Mitchell, 2000, Boer and Yu, 2003b) but not all (Watterson, 2000) climate models and it is important to understand the cause, magnitude, and implications of this evolution as it applies to the climate system.

In this context, the usual equilibrium $2xCO_2$ sensitivity is a broad, 1st order, and very visible measure of the response of the climate system to a given forcing which indirectly conveys some information on regional patterns of change and which is useful for calibrating simpler models. It is, however, a heavily averaged parameter and attention should be directed also toward the geographical pattern and temporal evolution of the feedback g and the normalized feedback parameter $\Lambda_f = g / \langle T' \rangle$ together with the associated effective global sensitivity $s(t)$ in order to more completely understand and characterize the behaviour of the system and its evolution.

R1. It is recommended that climate modelling groups calculate the temporal evolution of the global effective sensitivity $s(t)$ as part of their standard diagnostic information (including providing it as part of the IPCC data). The geographical pattern and evolution of the forcing f , feedback g and of the feedback parameter Λ_f , including its components, are all important and should be diagnosed and displayed.

The purpose includes documenting the evolution of sensitivity with climate state and providing information on the nature and magnitude of the feedbacks operating in climate models under the range of IPCC (and other) forcing scenarios.

2 Climate sensitivity and temperature change

Climate sensitivity and the evolution of temperature are closely related aspects of the climate system and its behaviour under global warming. For the globally averaged system in (3-4) and assuming that $\langle dh'/dt \rangle \approx \kappa \langle T' \rangle$, i.e. that the system is not in equilibrium and that rate of storage of heat in the deep ocean depends on the temperature change, then to first order

$$\delta \langle T' \rangle \approx \left(\frac{s}{1 + \kappa s} \right) \delta \langle f \rangle \text{ whence } \delta \langle T' \rangle_2 \approx \left(\frac{\delta \langle f \rangle_2}{\delta \langle f \rangle_1} \right) \delta \langle T' \rangle_1 \quad (5)$$

for s and κ constant or slowly varying. Thus the temperature change for a given forcing depends non-linearly on both the sensitivity and the rate of storage of heat in the ocean, but the temperature change in the future is related to the temperature change in the past by the ratio of the forcing changes.

3 Constraining climate sensitivity

Constraints on climate sensitivity based on the past behaviour of the climate system are of considerable interest to the IPCC. Constraints may consist of a range of possible $\langle T'_{2x} \rangle$ values as, for instance, the 1.5 to 4.5°C range of the IPCC reports or, preferably, the probability distribution $p(s)$ which reflects the uncertainty in a more quantitative way. Gregory et al. (2003) obtain an estimate of $p(s)$ by estimating the values of each of $\langle T' \rangle$, $\langle f \rangle$ and $\langle dh'/dt \rangle$ together with a measure of uncertainty from observations and other information. Taking these to represent means and standard deviation of normally distributed variables, the usual calculus of probabilities gives the probability distribution of a function of these variables namely of sensitivity s from (4).

The form of (4) implies that $p(s)$ will be asymmetric with a fairly sharp cut off for low values of s and a longer tail for larger values of s . This general structure is seen in a range of other studies (e.g. Forest et al., 2000; Andronova and Schlesinger, 2001; Forest et al., 2001; Harvey and Kaufman, 2002; Knutti et al., 2002; Forest et al., 2002; and Knutti et al., 2003) that infer $p(s)$ from the behaviour of models of the climate system constrained by the observational record. The general approach in these, and other studies, is to constrain the climate sensitivity, and other parameters including the radiative forcing, by fitting model results to observed changes. The 90% confidence interval from studies such as these typically encompasses the IPCC sensitivity range of 1.5 to 4.5°C.

The instrumental record thus provides the possibility of inferring constraints on climate sensitivity which depend, however, on the accurate estimation of the forcing and the successful attribution of observed warming among the several external forcings including GHGs, volcanoes and solar change. The uncertainty in the forcing and in the nature and magnitude of the natural

variability of the system together with different model behaviours lead to somewhat different estimates of s and of $p(s)$ and this leads to the following recommendation.

R2. Constrains on system sensitivity may be tightened by the study of particular climate perturbations where the forcing and the response are comparatively well known. It is recommended that IPCC coupled model experiments include the simulation of the climate perturbation associated with the Pinatubo eruption with special attention to the nature and timescale of the response, the role of ocean uptake, and the bounds that can be placed on climate sensitivity.

The same general approaches may be used to investigate climate sensitivity based on forced climate variations that occurred in periods preceding the instrumental observational record. Studies such as those of Hoffert and Covey (1992) and Hegerl et al. (submitted) take this approach. This requires that suitable and reliable non-instrumental, i.e. proxy, temperature and other information is available together with information on climate forcing. This need motivates the following suggestion.

R3. Studies of past forced climate variations directed toward constraining sensitivity should be pursued. This depends on the availability of forcing and response data sets which must be vigorously but carefully developed.

Finally, model behaviour indirectly gives information on the range of s and the distribution $p(s)$ in so far as the models are representative of the climate system. One approach is to consider the collection of model sensitivities from complex climate models as a random sample from the population of climate models embodying our current understanding of the climate system. These results, as embodied in Figure 9.18 of IPCC2001 for instance, may be used to estimate $p(s)$ under this assumption. A related approach, discussed in detail in Section 1, investigates $p(s)$ by perturbing model parameters.

4 *Constraining temperature evolution*

Climate sensitivity by itself does not determine the evolution of temperature for a given IPCC forcing scenario. One of the results of IPCC2001 (Chapter 9) was to show that many models reasonably successfully simulate the observed 20th century global warming even though climate sensitivities differ by a factor of 2 or more. The modelled temperatures subsequently diverge as the simulations proceed into the 21st century, however. Thus for early times, when T' is small, its evolution is initially determined by f and by ocean heat uptake and only secondarily by feedback strength/sensitivity.

For the next few decades, constraints on warming can likely be more directly inferred from the past climate record than from the sensitivity itself since from (5) future temperature change is related to past temperature change by the ratio of the changes in forcing. This contrasts with the inferred sensitivity which depends in a non-linear way on both ocean heat uptake and the observed temperature trend. Of course this depends on the comparative constancy of sensitivity and ocean heat uptake.

In subsequent decades, as warming increases, feedback/sensitivity becomes a controlling determinant of the simulated warming in models and the real system. Some constraints are nevertheless possible based on the past record as investigated, for instance by Allen et al. (2000, 2003), Stott and Kettleborough (2002), Sokolov et al., (2003), Knutti et al., (2003) among others.

The continued study of the physical processes determining both sensitivity and the temperature record is important. Particular attention should be paid to ocean heat uptake and those data sets that aid the specifications of past forcing and permit the attribution of the climate perturbations to these forcings.

R4. A vigorous program of extending and synthesizing the instrumental record should be undertaken. In particular, new variables should be analyzed and the temporal and spatial coverage of the usual variables such as temperature should be extended into the bodies of both the atmosphere and the ocean in a bid to constrain sensitivity and temperature.

References

- Allen, M., Wm. Ingram and D. Stainforth, 2002: Constraints on future changes in climate and the hydrologic cycle. *Nature*, **419**, 224-232.
- Allen, M. P. Stott, J. Mitchell, R. Schnur and T. Delworth, 2000: Quantifying the uncertainty in forecasts of anthropogenic climate change. *Nature*, **407**, 617-620.
- Andronova, N.G. and M.E. Schlesinger, 2001: Objective estimation of the probability density function for climate sensitivity. *J. Geophys. Res.*, **106**, 22605-22611.
- Boer, G.J. and B. Yu, 2003a: Climate sensitivity and response. *Clim. Dyn.*, **20**, 415-429.
- Boer, G.J. and B. Yu, 2003b: Climate sensitivity and climate state. *Clim. Dyn.*, **21**, 167-176.
- Forest, C.E, M.R. Allen, P.H. Stone, A.P. Sokolov, M.R. Allen and M.D. Webster, 2002: Quantifying uncertainties in climate system properties with the use of recent climate observations. *Science*, **295**, 113-117.
- Forest, C.E, M.R. Allen, A.P. Sokolov and P.H. Stone, 2001: Constraining climate model properties using optimal fingerprint detection methods. *Clim. Dyn.*, **18**, 277-295.
- Forest, C.E, M.R. Allen, P.H. Stone, and A.P. Sokolov, 2000: Constraining uncertainties in climate models using climate change detection techniques. *Geophys. Res. Lett.*, **27**, 569-572.
- Gregory, J.M., et al., 2002: An observationally based estimate of the climate sensitivity. *J. Clim.*, **15**, 3117-3121.
- Hansen, J., M. Sato, and R. Ruedy, 1997: Radiative forcing and climate response, *J. Geophys. Res.*, **102**(D6), 6831-6864.
- Harvey, L.D. and R.K. Kaufmann, 2002: Simultaneously constraining climate sensitivity and aerosol radiative forcing. *J. Clim.*, **15**, 2837-2861.
- Hegerl G. C., T. J. Crowley, and Wm. Hyde, 2004: Estimating climate sensitivity from paleo-data. To be submitted
- Hoffert, M.I. and C. Covey, 1992: Deriving global climate sensitivity from paleoclimate reconstructions. *Nature*, **260**, 573-576.

- Knutti, R., T. Stocker, F. Joos and G.-K. Plattner, 2002: Constraints on radiative forcing and future climate change from observations and climate model ensembles, *Nature*, **416**, 719-722.
- Knutti, R., T. Stocker, F. Joos and G.-K. Plattner, 2003: Probabilistic climate change projections using neural networks. *Clim. Dyn.*, **21**, 257-272.
- Murphy, J.M., 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. Clim.*, **8**, 57-80.
- Murphy J M. et al. (2004) Quantification of modelling uncertainties in a large ensemble of climate change simulations. *Nature*, **430**, 768-772.
- Senior, C.A. and J. Mitchell, 2000: The time-dependence of climate sensitivity. *Geophys. Res. Lett.*, **27**, 2685-2689.
- Senior, C.A., Wielicki, B.A., McAvaney, B. and Boer, G. J., Report on the joint WCRP CFMIP/ IPCC expert meeting on 'Climate Sensitivity and Feedbacks'. Briefing material for Paris meeting
- Sokolov, P, C.E. Forest, and P. H. Stone, 2003: Comparing oceanic heat uptake in AOGCM transient climate change experiments. *J. Clim.*, **15**, 1573-1582.
- Stainforth, D. A. et al. (2004). Evaluating uncertainty in the climate response to changing levels of greenhouse gases. Submitted to *Nature*.
- Stott, P. and J. Kettleborough, 2002: Origins and estimates of uncertainty in predictions of twenty-first century temperature rise, *Nature*, **416**, 723-726.
- Watterson, I.B., 2000: Interpretation of simulated global warming using a simple model. *J. Clim.*, **13**, 202-215.

