

## Does the Last Glacial Maximum constrain climate sensitivity?

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[1] Four simulations with atmosphere-ocean climate models have been produced using identical Last Glacial Maximum ice sheets, topography and greenhouse gas concentrations. Compared to the pre-industrial, the diagnosed radiative feedback parameter ranges between  $-1.30$  and  $-1.18 \text{ Wm}^{-2}\text{K}^{-1}$ , the tropical ocean sea-surface temperature decreases between  $1.7$  and  $2.7^\circ\text{C}$ , and Antarctic surface air temperature decreases by  $7$  to  $11^\circ\text{C}$ . These values are all compatible with observational estimates, except for a tendency to underestimate the tropical ocean cooling. On the other hand, the same models have a climate sensitivity to  $\text{CO}_2$  concentration doubling ranging between  $2.1$  and  $3.9 \text{ K}$ . It is therefore inappropriate to simply scale an observational estimate of LGM temperature to predict climate sensitivity. This is mainly a consequence of the non-linear character of the cloud (mainly shortwave) feedback at low latitudes. Changes in albedo and cloud cover at mid and high latitudes are also important, but less so. **Citation:** Crucifix, M. (2006), Does the Last Glacial Maximum constrain climate sensitivity?, *Geophys. Res. Lett.*, 33, L18701, doi:10.1029/2006GL027137.

### 1. Introduction

[2] The response of the climate system to the increase in greenhouse gas concentrations is commonly characterised by the *climate sensitivity*, that we denote  $\Delta T_{2\times\text{CO}_2}$ , and which is defined as the global mean surface temperature response to a doubling of  $\text{CO}_2$  atmospheric concentration once the system has reached an equilibrium state [e.g., *Schlesinger and Mitchell*, 1987]. Estimates of climate sensitivity of most general circulation models vary among models between  $1.5$  and  $4.5 \text{ K}$  [*Cubasch et al.*, 2001], but it exceeds  $6 \text{ K}$  in certain models [*Webb et al.*, 2006].

[3] Climate sensitivity may formally be expressed by considering the energy balance of the global system:

$$C \frac{d\Delta T}{dt} = Q \quad (1)$$

where  $C$  is the heat capacity of the system and  $Q$  is the net downward radiative flux at the top of the atmosphere.  $Q$  is then decomposed in  $F + R$  where  $F$  is the *forcing* (the radiative perturbation imposed to the system) and  $R$ , the response. The *climate feedback parameter*  $\lambda$  is defined as  $\lambda = R/\Delta T$ . Thus:

$$\Delta T = \frac{Q - F}{\lambda} \quad (2)$$

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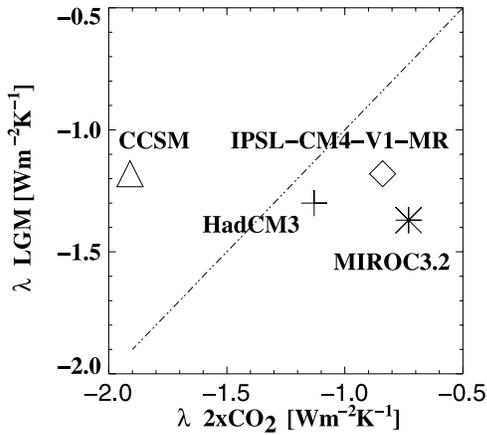
[4] At equilibrium,  $Q = 0$ , so that  $\Delta T_{2\times\text{CO}_2} = -F_{2\times\text{CO}_2}/\lambda_{2\times\text{CO}_2}$ . The climate feedback parameter may in principle be defined for any type of forcing and the system does not need to be in radiative equilibrium. In particular, it may be diagnosed using observations of recent climate change [*Andronova and Schlesinger*, 2001; *Gregory et al.*, 2002; *Annan and Hargreaves*, 2006].

[5] However,  $\lambda$  generally depends on the nature of the forcing, its amplitude and the way it is defined (which is partly arbitrary). *Hansen et al.* [2005] diagnosed  $\lambda$  for various forcings (greenhouse gases, stratospheric water vapour, ozone, land-use change) and different definitions of the forcing using the GISS model E. Most of the forcings relevant to recent and future climate change yield values of  $\lambda$  similar to  $\lambda_{2\times\text{CO}_2}$ , plus or minus 30%.

[6] Here, we focus on the Last Glacial Maximum climate (21,000 years ago). Observational estimates of the surface temperature change between that period and today exist for Antarctica ( $-9 \pm 2^\circ\text{C}$  [*Jouzel et al.*, 2003]), Greenland ( $-20 \pm 2^\circ\text{C}$ ) [*Cuffey and Clow*, 1997; *Dahl-Jensen et al.*, 1998] and the tropical ocean ( $-2.7 \pm 0.5^\circ\text{C}$  [*Ballantyne et al.*, 2005; *Lea*, 2005]). Inverse ice sheet modelling provides a further estimate of the cooling ( $4$  to  $10^\circ\text{C}$ ) over the northern hemisphere continents [*Bintanja et al.*, 2005]. The global temperature change is therefore estimated to be comprised between  $3$  and  $9^\circ\text{C}$  with 95% confidence [*Annan and Hargreaves*, 2006].

[7] Assuming that the system was in radiative equilibrium at the Last Glacial Maximum, it may be considered that the cooling was caused by different radiative forcings: (i) an albedo forcing, related to the presence of large ice sheets absent today, the change in sea-level and vegetation changes (around  $-4 \text{ Wm}^{-2}$ , this study) (ii) a reduction of greenhouse gas concentrations (about  $-2.85 \text{ Wm}^{-2}$ ) (iii) other forcings, difficult to quantify, such as increased dust concentration (about  $-1 \text{ Wm}^{-2}$  [*Claquin et al.*, 2003]). There is also a small contribution due to the surface being, on average, more elevated than today. The ratio between the radiative response and temperature response at the LGM is thus of the order of  $-1.3 \text{ Wm}^{-2}\text{K}^{-1}$ . Accounting for uncertainties, *Annan and Hargreaves* [2006] provide 95% confidence limits of  $-0.8$  and  $-2.9$ , respectively (cf. *Hansen et al.* [1984] and *Hoffert and Covey* [1992] for previous estimates). Another method to estimate  $\lambda$  from Quaternary climate change consists in analysing the variance in temperature explained by greenhouse gases in palaeoclimate records covering several 100,000 years [*Lorius et al.*, 1990; *Lea*, 2005]. These analyses yield values comprised between  $-2.5$  and  $-1.5 \text{ Wm}^{-2}\text{K}^{-1}$ .

[8] Are these estimates a good indicator of  $\lambda_{2\times\text{CO}_2}$ ? *Annan et al.* [2005] observe a reasonably good correlation between tropical LGM cooling and climate sensitivity in an ensemble of experiments with MIROC3.2 with different internal parameters. An even better correlation is found in



**Figure 1.** Radiative feedback parameters diagnosed in response to LGM forcing vs  $\text{CO}_2$  doubling in four atmosphere-ocean climate models.

CLIMBER2 [Schneider von Deimling et al., 2006]. However, the relationship between  $\lambda_{2\times\text{CO}_2}$  and  $\lambda_{2\times\text{LGM}}$  may vary among models that differ structurally. For example,  $\lambda_{\text{LGM}} = 0.85\lambda_{2\times\text{CO}_2}$  in the atmosphere-slab model HadSM2 [Hewitt and Mitchell, 1997], i.e., the surface temperature increases less in response to  $\text{CO}_2$  doubling than expected from the LGM experiment (also found in work by Broccoli and Manabe [1987] and Kim [2004]). This effect is commonly attributed to changes in radiative forcing being more effective near the surface and in high latitudes [Hansen et al., 1997] and/or sea-ice and snow albedo feedback becoming more important at lower temperatures. Contradictory to this, the GISS model E is slightly more sensitive to an increase in  $\text{CO}_2$  concentration than to a decrease [Hansen et al., 2005]. Furthermore, it is noted that the relative contribution of greenhouse gases to the total tropical cooling at the LGM is estimated to be minor [Hewitt and Mitchell, 1997], medium [Shin et al., 2003] or large [Kim, 2004] depending on the climate model.

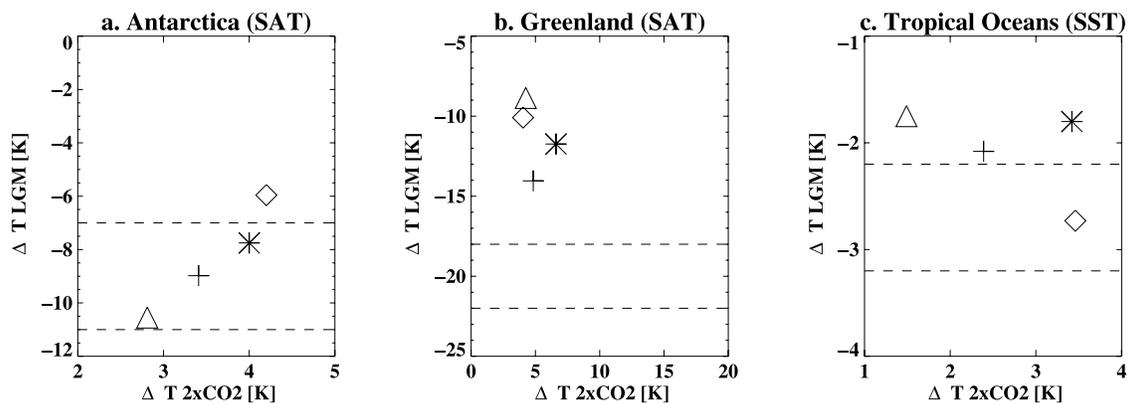
[9] Definitive conclusions have not been drawn so far due to significant differences in the above cited experimental designs. Intercomparison projects now allow us to

compare predictions of  $\lambda_{2\times\text{CO}_2}$  and  $\lambda_{2\times\text{LGM}}$  with greater rigor. This is the goal of the present study.

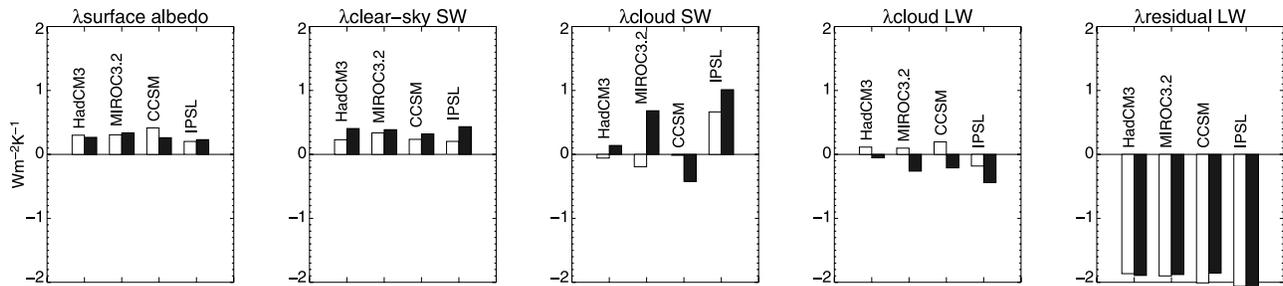
## 2. Methods

[10] Last Glacial Maximum experiments and corresponding pre-industrial control experiments were carried out with atmosphere-ocean models in the framework of the second phase of the Palaeoclimate Modelling Intercomparison Project (PMIP2 [Harrison et al., 2002; Crucifix et al., 2005]). Output from four models are available: HadCM3, NCAR-CCSM3.0 (T42), IPSL-CM4 and MIROC3.2 (medium resolution). Models are configured using the same surface orography, land sea-mask, orbital parameters and greenhouse gas concentrations (see <http://www-lsce.cea.fr/pmip2> for model descriptions and experimental details). Experiments are run for several centuries, until radiative equilibrium is close, and diagnostics are averaged over 100 years. The radiative forcing is made of two contributions: the albedo forcing and the greenhouse gas forcing. The albedo forcing is estimated by approximation (K. E. Taylor et al., Estimating shortwave radiative forcing and response in climate models, submitted to Journal of Climate, 2006) (hereinafter referred to as Taylor et al., submitted manuscript, 2006) to the partial radiative perturbation method [Wetherald and Manabe, 1988]. The approximation was tested in two climate models (Taylor et al., submitted manuscript, 2006) and only requires output available in the PMIP2 database. The greenhouse gas forcing is estimated as  $F_{2\times\text{CO}_2}(-2.85/3.70)$ .  $F_{2\times\text{CO}_2}$  is the stratosphere-adjusted estimate of the  $\text{CO}_2$  doubling forcing at the tropopause (shortwave + longwave) provided by the different modelling groups;  $-2.85 \text{ Wm}^{-2}$  and  $+3.70 \text{ Wm}^{-2}$  are the greenhouse gas radiative forcings of the LGM and  $\text{CO}_2$  doubling, respectively, estimated as in [Intergovernmental Panel on Climate Change, 2001, p. 358].

[11] A 600-year long  $\text{CO}_2$  doubling experiment of CCSM3.0 (T42) is available with the corresponding control from the earth system grid portal (<http://www.earthsystemgrid.org/>).  $\text{CO}_2$  doubling experiments (1% per year increase and then stabilisation) for the other models are available from the IPCC AR4 database ([http://www.pcmdi.llnl.gov/ipcc/about\\_ipcc.php](http://www.pcmdi.llnl.gov/ipcc/about_ipcc.php)) but they are typically 200-year long. Averages over the last 50 years are retained.



**Figure 2.** Change in (a) Antarctic surface air temperature (b) Greenland surface air temperature and (c) sub-tropical ( $30^\circ\text{S}$ ;  $30^\circ\text{N}$ ) ocean sea-surface temperature simulated by the climate models in response to LGM and  $\text{CO}_2$  doubling forcings. Horizontal dashed lines are observational estimates for the LGM (confidence interval of 66%).



**Figure 3.** Partial radiative feedback parameters to the LGM forcing (white) and CO<sub>2</sub> doubling (dark).

Temperatures shown are further scaled by  $F/(F - Q)$  to account for any remaining radiative imbalance, which is equivalent to assuming that  $\lambda$  has reached a constant level (reasonable according to *Senior and Mitchell* [2000], and as verified using CCSM output).

### 3. Results

[12] Figure 1 shows  $\lambda_{\text{LGM}}$  vs.  $\lambda_{2 \times \text{CO}_2}$  diagnosed in the four models. All models have a  $\lambda_{\text{LGM}} \simeq -1.3 \text{ Wm}^{-2}\text{K}^{-1}$ , which is very close to the best guess observational estimate quoted above. On the other hand, these models predict  $\lambda_{2 \times \text{CO}_2}$  between  $-1.9$  and  $-0.8 \text{ Wm}^{-2}\text{K}^{-1}$ . This implies that CCSM warms less in response to a CO<sub>2</sub> doubling than expected from the LGM experiment, while MIROC3.2 warms more. Therefore, the sole observational constraint on  $\lambda_{\text{LGM}}$  does not allow us to exclude any climate sensitivity within this range on the basis of the four models analysed. Models are consistent with the estimates of temperature change over Antarctica (6 to 11 K cooling) and over the tropical ocean (1.7 to 2.7 K) (Figure 2). A slight underestimate of the tropical ocean cooling is actually expected given the omission of some forcings in the experimental design. The same models underestimate by several degrees the Greenland cooling. The issue has already been commented on [Masson *et al.*, 2006] and is not further discussed here. More relevant to the present purpose is that the ratio between LGM cooling and CO<sub>2</sub> doubling warming varies by a factor of two from one model to the next: the model showing the largest CO<sub>2</sub> doubling warming (MIROC3.2) has the smallest LGM cooling over Antarctica. Likewise, MIROC3.2 is one of the two models showing the smallest tropical cooling at the LGM, while it is one of the two models showing the largest temperature increase in this area in response to CO<sub>2</sub> doubling. It is therefore inappropriate to simply scale an observational estimate of global LGM temperature to predict climate sensitivity.

[13] The discrepancy between Antarctic cooling at the LGM and warming in response to CO<sub>2</sub> doubling is partly a consequence of Antarctica being, at the LGM, about 250 m higher than today. The altitude difference is the same in all models, but the contribution of this orographic forcing to the temperature decrease is model-dependent. This is, however, only part of the explanation.

[14] Elsewhere, the reasons for the asymmetric character of the responses to LGM vs CO<sub>2</sub> doubling can be better understood by decomposing the radiative response  $R$  into the contributions due to the change in surface albedo ( $R_{\text{alb}}$ ),

change in cloud cover and cloud optical properties in the shortwave ( $R_{c,sw}$ ) and longwave ( $R_{c,lw}$ ), in clear-sky shortwave optical properties ( $R_{cs,sw}$ ), and finally a long-wave residual ( $R_{lwr}$ ), that includes the contributions of the changes in surface temperature, lapse-rate and water-vapour concentration.

[15]  $R_{\text{alb}}$ ,  $R_{c,sw}$  and  $R_{cs,sw}$  are diagnosed by approximate partial radiative perturbation (Taylor *et al.*, submitted manuscript, 2006).  $R_{c,lw}$  is simply estimated as the change in longwave cloud forcing [Cess *et al.*, 1990]. The values are divided by the global mean temperature change to form a series of partial feedback parameters, denoted  $\lambda_{\text{alb}}$ , etc. Global averages for the four models and the two experiments (LGM and  $2 \times \text{CO}_2$ ) are plotted on Figure 3. Spatial distributions are shown in the auxiliary material<sup>1</sup>.

[16] Surface albedo is generally a small contributor to the difference in the global responses to LGM and CO<sub>2</sub> doubling. Only in CCSM is the difference notable ( $\lambda_{\text{alb}}$  is  $0.15 \text{ Wm}^{-2}\text{K}^{-1}$  larger in the LGM experiment) due to a greater response of southern sea-ice in the LGM experiment, although partly compensated for by a greater response of Arctic sea-ice to CO<sub>2</sub> doubling. Furthermore,  $\lambda_{\text{alb}}$  is larger in response to a  $2 \times \text{CO}_2$  forcing than to the LGM forcing in 2 models, and smaller in the two others. This therefore contradicts the argument that the albedo feedback is responsible for models being systematically more sensitive as climate is colder.

[17] The clear-sky atmospheric shortwave feedback is mainly due to an increase in water-vapour concentration. It is positive in all models, and slightly larger in the CO<sub>2</sub> doubling experiment. This is consistent with the Clausius-Clapeyron relation, and contributes to make all the models slightly more sensitive to CO<sub>2</sub> doubling.

[18] Changes in the shortwave atmospheric cloud feedback explain a large fraction of the difference between models, and between experiments. In MIROC3.2, there are two main reasons for the greater shortwave cloud feedback in response to CO<sub>2</sub> doubling compared to LGM: (i) a larger response (decrease with warming) of low-level cloud cover and/or optical thickness over subtropical anticyclones in the CO<sub>2</sub> doubling experiment. This explains most of the difference in climate sensitivity between LGM and CO<sub>2</sub> doubling in this model; (ii) the LGM experiment shows a decrease in mid-level cloud along the northern summer intertropical convergence zone, especially in the Northern Pacific. A similar feedback occurs in response to

<sup>1</sup>Auxiliary materials are available in the HTML. doi:10.1029/2006GL027137.

CO<sub>2</sub> doubling, but it is weaker. This tends to make the model more responsive to CO<sub>2</sub> doubling. However, this effect is to a good extent compensated for in the longwave (Figure 3 shows that the longwave cloud feedback is more negative in response to CO<sub>2</sub> doubling). The cloud response in CCSM is particularly complex and difficult to analyse. The CO<sub>2</sub> doubling experiment is characterised by a large number of grid points located at low latitudes having a negative shortwave feedback and a small or negative longwave feedback. This suggests more or thicker low or mid-level clouds combined with less or lower high thin clouds [Webb *et al.*, 2006]. These cloud feedbacks are much weaker in the LGM experiment. HadCM3 and IPSL show more similar responses to LGM and CO<sub>2</sub> doubling. IPSL simulates a slightly larger positive feedback in the CO<sub>2</sub> doubling experiment due to the response of low and/or mid-level cloud over the southern ocean, North of 45°S. HadCM3 shows a greater response of equatorial convective clouds around the western Pacific to the CO<sub>2</sub> forcing compared to the LGM, but again the impact on the global radiative balance is small due to cancelling shortwave and longwave cloud feedbacks in this region.

[19] In addition to the different responses of low and mid-latitude clouds, there is, in all models, a negative shortwave cloud feedback specific to the LGM coming from the suppression of cloud over the ice sheets. The resulting difference in  $\lambda_{c,sw}$  (contribution of area North of 40°N to the global LGM vs.  $2 \times$  CO<sub>2</sub> difference) is comprised between  $-0.11$  (IPSL) and  $-0.27 \text{ Wm}^{-2}\text{K}^{-1}$  (MIROC3.2). This implies that this effect is accounted for similarly in the different models, and therefore does not contribute much to the inter-model difference.

[20] The longwave residual is large and negative in all models. It is primarily related to the change in surface longwave emission associated with the temperature change, amplified by the water-vapour feedback. It is of the order of  $-1.9 \text{ Wm}^{-2}\text{K}^{-1}$  whatever the model and forcing.

#### 4. Discussion and Conclusion

[21] The four models analysed here form a small sample of current state-of-the-art models used to predict the future climate, though they are fairly representative in the sense that they almost cover the commonly accepted range of likely climate sensitivity. It is therefore unexpected that these models have similar global mean radiative responses to the Last Glacial Maximum Climate. It was also shown that the ratio between LGM cooling and  $2 \times$  CO<sub>2</sub> warming depends greatly on the model. Therefore, climate sensitivity cannot be easily estimated from the Last Glacial Maximum global temperature.

[22] The LGM global radiative responses being similar across models may be an artifact of the small sample (four models). However, the reasons for LGM and CO<sub>2</sub> feedback factors differing so much have been identified and they appear reasonable. The main one is that subtropical shallow-convective clouds do not respond linearly to temperature change. This particular effect is probably not a direct consequence of the presence of the ice sheets on atmosphere dynamics, and therefore emphasises the fundamentally non-linear nature of the climate response to the radiative forcing. This result has two consequences. First, it certainly encour-

ages a more systematic analysis of the dependency of the feedback factor on the nature and the sign of the forcing in climate models. Second, global estimates of the LGM temperature only weakly constrain climate sensitivity for two reasons: (i) the forcing is not known accurately and (ii) the ratio between LGM and CO<sub>2</sub> feedback factors cannot be accurately estimated from current state-of-the-art coupled models. This implies that careful model-data comparisons on the details of the spatial distribution of changes in temperature and precipitation at the LGM are needed to identify the “best models”, that is, those that reliably predict the response of climate dynamics to a given forcing. Global temperature is not sufficient.

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