In the first section of this Supplement describe the different parameter sets that are used in the main paper. In the second section, four types of experiments are presented to characterise the model behaviour with each of these parameter sets. The purpose of this section is to establish the framework in which the subsequent simulations are carried out. In the third section, a last simulation of the climate evolution over the last millennium is described. It serves as a basis for the simulations over the last century that are discussed in the main paper.

#### 1. The parameter sets

#### a. Climatic parameter sets

Several physical parameters of the model may significantly impact the model response to an external perturbation. Although other studies (Collins et al., 2007; Brierley et al., 2010) suggested that the impacts of ocean parameter uncertainty on climate response are small compared to perturbed atmosphere parameters, the parameters are selected in both the atmospheric (ECBilt) and oceanic (CLIO) components of the model. We performed more than one hundred simulations of the present-day climate modifying the nominal value of one or several parameters and we kept only combinations (parameter sets) that (1) provide a realistic present-day simulation and (2) cover the best the phase space (climate sensitivity vs MOC sensitivity), leading to contrasted responses to a doubling of CO<sub>2</sub> concentration (climate sensitivity) and to an additional freshwater flux in the North Atlantic (MOC sensitivity) (Section 2). The parameter values are chosen within their range of uncertainty. In doing so, we want to keep a grip on the parameter sets in order to be able to identify the impact of each parameter set on the simulated climate. This is also why the number of parameter sets included in this study is relatively small. Table I gives the values for the nine selected climatic parameter sets and a short description of them (see also Goosse et al.; 2007). The Rayleigh damping term of the equation of the quasi-geostrophic potential vorticity includes two parameters. The first one,  $\lambda 2$ , corresponds to the 300-500hPa layer of the model, while the second one,  $\lambda 4$ , corresponds the 500-800 hPa layer (see equation 1 of Opsteegh et al. (1998) and equation 11 of Haarsma et al. (1996)).

The simple longwave radiative scheme of LOVECLIM is based on an approach termed the Green's function method (Chou and Neelin, 1996; Schaeffer et al., 1998).

The scheme could be briefly represented for clear-sky conditions by the following formula for all the model levels:

Flw= Fref + FG(T',GHG')+ G1 \* amplw\*(q')\*\*explw

where Flw is the longwave radiation, Fref a reference value of the radiation when temperature, humidity and the concentration of greenhouse gases are equal to the reference values, and FG a function, not explicitly described here, allowing to compute the contribution associated with the anomalies compared to this reference in the vertical profile of temperature (T') and in the concentrations of the various greenhouse gases in the atmosphere (GHG'). The last term represents the anomaly in the longwave radiation due to the anomaly in humidity q'. The coefficients Fref, G1 and those included in the function FG are spatially dependent. All the terms have been calibrated to follow as closely as possible a complex GCM longwave radiative scheme (Schaeffer et al., 1998), but large uncertainties are of course related to this parameterisation, in particular as the model only computes one mean relative humidity between the surface and 500 hPa, the atmosphere above 500 hPa being supposed to be completely dry. Here, both amplw and explw are varied with a clear impact on the influence of humidity changes on longwave fluxes.

Clouds are probably one of the most important components, and yet poorly known, of the climate system. In LOVECLIM, cloudiness is prescribed according to present-day climatology. Although it is a potentially important source of uncertainty, we are unable to investigate it because of the crude representation of clouds in the model and of it simplified atmospheric component. However, we manage to compensate for this weakness through other parameters such as the representation of the water vapour effect in the radiative scheme.

The albedo of the ocean in LOVECLIM depends on the season and location. At each time step, it is multiplied by albcoef. For a typical albedo of the ocean of 0.06, using a value of 1.05 for albcoef increases the value of the albedo to 0.063.

The albedo of sea ice (albice) is computed by the scheme of Shine and Henderson-Sellers (1985), which uses different values for the albedo of snow, melting snow, bare ice and melting ice. For thin ice, the albedo is also dependent on the ice thickness. If albice is different from zero in the experiments discussed here, the value of the albedo in the model is increased by albice for all the snow and ice types. The minimum vertical diffusion coefficient in the ocean follows a vertical profile similar to the one proposed by Bryan and Lewis (1979), as explained in detail in Goosse et al. (1999). The coefficient avkb is a scaling factor that multiplies the minimum value of the vertical diffusivity at all depths. A value of avkb of 1 (1.5, 2, 2.5) corresponds to a minimum background vertical diffusivity in the thermocline of  $10^{-5}$  m<sup>2</sup>/s ( $1.5 \times 10^{-5}$ ,  $2.0 \times 10^{-5}$ ,  $2.5 \times 10^{-5}$  m<sup>2</sup>/s).

The semi-implicit numerical scheme used for the Coriolis term in the barotropic and baroclinic horizontal momentum equation in LOVECLIM1.0 (Driesschaert et al., 2007) induced too much numerical noise. Therefore, in LOVECLIM1.1, it is computed in a totally implicit way for all the simulations, except those using climatic parameter set 11. The former scheme is kept here in order to provide an easier comparison with the results of LOVECLIM1.0. Because of the larger implicit diffusion associated with this scheme, a lower value of the explicit diffusion is applied with climatic parameter set 11.

As ECBilt systematically overestimates precipitation over the Atlantic and Arctic Oceans, it has been necessary to artificially reduce the precipitation rate over the Atlantic and Arctic basins (defined here as the oceanic area north of 68°N). The corresponding water is dumped into the North Pacific, a region where the model precipitation is too weak (Goosse et al., 2001). CorA corresponds to the percentage of reduction of the precipitation in the Atlantic.

# b. The carbon cycle parameter sets

In addition to the nine climatic parameter sets (based on the parameters described in the previous section), we define three parameter sets inducing different responses of the carbon cycle model (i.e. carbon cycle parameter sets; Table II). The key parameters for the carbon cycle are chosen among those that have a strong impact on the marine biogeochemical cycle and on the response of the atmospheric  $CO_2$  to an emission scenario. More precisely, they deal with the continental vegetation fertilization effect, on the one hand, and the rain ratio and the vertical flux of particulate organic matter (POM) in the ocean, on the other hand (Mouchet, 2010).

The fertilization effect constitutes a negative feedback on  $CO_2$ . The effect of  $CO_2$  on continental vegetation uptake is parameterised with the following formula:

NPP=NPP0 (1+ $\beta$  ln (pCO<sub>2</sub>/pCO<sub>2ref</sub>)), where NPP is the net primary production, pCO<sub>2</sub> the CO<sub>2</sub> atmospheric pressure, and NPP0 and pCO<sub>2ref</sub> the NPP and pCO<sub>2</sub> for a reference state, respectively. As there are different fertilization responses according to the ecosystem (e.g. Houghton et al., 2001), we separated the fertilization effect in two terms: one for grass ( $\beta$ g) and one for forests ( $\beta$ t).

The vertical flux of POM is one factor controlling the sequestration of CO<sub>2</sub> in the deep ocean. It is represented in the model by a power law  $z^{\alpha}$ , with z the depth (Martin et al., 1987). The  $\alpha$  factor exhibits a large range of values (e.g. Martin et al., 1987; Suess, 1980; Berger et al., 1987; Betzer et al., 1984). Such a range could be explained by differences in ecosystems (e.g. Klaas and Archer, 2002). Hence carbon cycle parameter set 1 considers different profiles for diatoms ( $\alpha_{diatom}$ ) and other species ( $\alpha_{others}$ ) in order to account for the sensitivity of  $\alpha$  on ecosystem composition.

The buildup of calcium carbonate shells in the surface ocean results in a CO<sub>2</sub> source to the atmosphere, while dissolution constitutes a sink. The  $\Psi_{zoo}$  parameter represents the contribution of zooplankton in the precipitation of biogenic CaCO<sub>3</sub>. A larger value of this parameter implies a larger rain ratio. The rain ratio is defined as the ratio of inorganic carbon content over that of organic carbon in biogenic particles sinking to depth.

# 2. Preliminary experiments

# a. The pre-industrial climate

An equilibrium experiment (prefix E; Table III), under pre-industrial conditions, is performed using each of the nine selected climatic parameter sets to ensure that all the selected parameter sets yield a reasonable pre-industrial climate. Carbon cycle parameter set 2 used here corresponds to the parameter values used in previous studies (e.g., Menviel et al., 2008a; Plattner et al., 2008). The various forcings are kept constant. No volcanic eruption is considered. The greenhouse gas concentrations are kept to their 1750 values (all years are in AD). Simulated climatic variables are compared to recent observations. Pre-industrial and present-day climates are slightly different. Here, we do not want to demonstrate that the simulated pre-industrial equilibrium climate reproduces perfectly the real pre-industrial climate. Rather, we show first that all the parameter sets lead to similar pre-industrial climates.

states. The difference between recent and pre-industrial climates is thus of minor importance in the present study. The model-data comparison mostly put forward the systematic biases of the model that are present with almost all parameter sets (e.g. Figure 1) as well as in other versions of the model (Goosse et al., 2001, 2010).

Table IV displays some global features simulated by LOVECLIM using the different parameter sets under pre-industrial forcing. For all the parameter sets, the globally averaged annual mean surface temperature is slightly too high, varying between 15.8 and 16.4°C, the main overestimation being observed at low latitudes (Figure 1a). LOVECLIM underestimates precipitation in the equatorial region (Figure 1b). This model feature is a consequence of the quasi-geostrophic approximation, which induces difficulties to simulate a correct Hadley cell (Renssen et al., 2002); it is not significantly modified by any parameter set. Mid- and high latitude precipitation is more properly represented than equatorial precipitation, independent of the parameter set. The model overestimates the tree fraction (Figure 1c) at all latitudes, whatever the parameter set. This overestimation is mostly at the expanses of the grass fraction, except in the mid-to-high southern latitudes (southern South America), where the cold desert area is underestimated. This general feature of the model is related with the overestimation of temperature over land, which is in favour of tree growth. The maximum of the strength of North Atlantic MOC (i.e. the annual mean value of the maximum of the North Atlantic meridional overturning streamfunction below the Ekman layer) varies between 17 and 28 Sv. These values lie within the range given by GCMs (e.g. Dixon and Lanzante, 1999; Gent, 2001). The sea ice extent in the Northern Hemisphere varies seasonally from a maximum between 14.3 and 15.1×10<sup>6</sup>km<sup>2</sup> in March to a minimum between 6.7 and 9.3×10<sup>6</sup>km<sup>2</sup> in September, while observations (average value between 1979 and 2000; Comiso and Nishio, 2008) give a maximum of less than  $16 \times 10^{6}$  km<sup>2</sup> and a minimum of  $6.9 \times 10^{6}$  km<sup>2</sup>.

# b. Sensitivity to CO<sub>2</sub> concentration

A first sensitivity experiment (prefix E, suffix 2CO, Table III) is performed starting from the equilibrium state described in the previous section. The atmospheric  $CO_2$  concentration is enhanced by 1% per year from the pre-industrial value until a doubled value is reached, i.e. after 70 years. It is held constant thereafter (Figure 2, left). This experiment provides a clear and strong climate signal as well as a good insight into the response of the atmosphere under perturbed conditions. Furthermore,

the response of LOVECLIM can be compared to the one of other models in similar conditions (e.g. Brovkin et al., 2006). The increase in global annual mean surface temperature after 1000 years in this sensitivity experiment is chosen as an index to characterise the response of the model to the prescribed perturbation (climate sensitivity). Carbon cycle parameter set 2 is used here.

The global annual mean surface temperature increase for the 9 climatic parameter sets ranges from 1.6 to 3.8°C after 1000 years (Table IV). Table IV also provides the temperature increase after 70 years in the two times CO<sub>2</sub> scenario (i.e. the transient temperature response or TCR), the effective climate sensitivity (Ceff) computed according to Gregory et al. (2002) and the equilibrium climate sensitivity (Equi). The temperature increase after 1000 years in our sensitivity experiment (CS) is already very close to the value of the effective climate sensitivity and the equilibrium climate sensitivity for the less sensitive parameter sets (112, 122, 212 and 222). Our parameters sets cover the likely range of climate sensitivity suggested by the IPCC (2007), i.e. 2.1°C to 4.4°C, based on GCM studies. It must be mentioned that, although LOVECLIM using climate parameter set 11 is not exactly the same as LOVECLIM1.0 used in Driesschaert et al. (2007), it shares many climatic features with this former version. In particular, its equilibrium sensitivity is rather low, i.e. 1.6°C. Figure 2 (right) displays the temperature evolution during the first 1000 years of the experiment. The rate of change is largest over the first 70 years of the simulation, when atmospheric CO<sub>2</sub> concentration is increasing.

The name of the experiments (e.g. first column in Table I and Table IV) has been designed to provide a quick overview of their main characteristics. Indeed, the first digit is related to the climate sensitivity. Its value goes from 1 to 5, corresponding to an increase less than 2.0°C to more than 3.5°C (by step of 0.5°C) in global annual mean surface temperature after 1000 years in the experiment described above (Figure 2).

#### c. Sensitivity to water hosing

In a second sensitivity experiment (prefix E, suffix HYS, Table IV) freshwater is added in the North Atlantic ( $20^{\circ}-50^{\circ}N$ ) with a linearly increasing rate of  $2\times10^{-4}$  Sv/yr during 1500 years. This results in a freshwater perturbation of 0.1 Sv after 500 years, 0.2 Sv after 1000 years and 0.3 Sv after 1500 years (Figure 3, left). This simulation, which allows assessing the stability of the North Atlantic MOC, provides a good

insight into the response of the ocean under perturbed conditions and can be compared with simulations performed with other models in similar conditions (e.g. Rahmstorf et al., 2005; Weber et al., 2007). Carbon cycle parameter set 2 is used here. The percentage of decrease of the maximum value of the meridional overturning streamfunction below the Ekman layer in the Atlantic Ocean after 1000 years in this water hosing experiment (at the time the perturbation reaches 0.2Sv) is chosen to characterise the response of the model to this perturbation (MOC sensitivity). LOVECLIM with parameter set 112, i.e. the closest LOVECLIM 1.0 used in Driesschaert et al. (2007), simulates a 20% reduction in the meridional overturning streamfunction after 1000 years. This decrease ranges from 17 to 62% for the other parameter sets (Table IV).

The MOC sensitivity is reflected in the second digit of the name of the experiments, through. It is one or two according to whether the decrease in the meridional overturning streamfunction after 1000 years is smaller or larger than 50%.

It is worth remembering that the model parameter sets lead to different pre-industrial equilibrium states with respect to the MOC (Table IV). Moreover, the time evolution of the meridional overturning streamfunction during the water hosing experiment shows several different patterns according to the model parameter sets (Figure 3, right). Indeed, for some climate parameter sets (e.g., 11), the meridional overturning streamfunction decreases almost linearly, while for others (e.g., 12), the MOC experiences a more abrupt weakening.

Figure 1 in the main paper shows that the phase space (MOC sensitivity vs. climate sensitivity) of our set of experiments is rather homogeneously covered as required by our initial objective. For comparison, the GCMs used in the IPCC-AR4 (Randall et al., 2007) have an equilibrium climate sensitivity ranging from 2.1°C to 4.4°C, with a mean value of 3.2°C. Although our parameter sets do not cover the full range of climate response to CO<sub>2</sub> increase and freshwater flux, they widely cover the range suggested by other studies (Randall et al., 2007). In an intercomparison of EMICs, Rahmstorf et al. (2005) showed that the width of the hysteresis curve, corresponding to changes in freshwater input, varies between 0.2 and 0.5 Sv. Amongst the models used in this intercomparison, those with three-dimensional ocean models (including ECBilt-CLIO, a former version of LOVECLIM, with general features similar to those of climatic parameter set 11) display a sharp weakening of the North Atlantic MOC for a

freshwater input of less than 0.3Sv. In our sensitivity experiment, which uses a slightly different setup, the meridional overturning streamfunction displays a very strong reduction for freshwater input ranging from 0.2Sv to 0.4Sv.

#### d. Sensitivity of the carbon cycle

We assess the sensitivity of the atmospheric CO<sub>2</sub> level to the carbon cycle parameter sets by performing a prognostic CO<sub>2</sub> experiment (prefix E, suffix TRA, Table III) for each of these sets. This transient simulation starts from an equilibrium state corresponding to the conditions prevailing in 1750. It runs until 3000 and is constrained by changes in the Earth orbital parameters (Berger, 1978) and in concentrations of greenhouse gases (GHGs) except CO<sub>2</sub>. In addition, the model is forced by anthropogenic emissions of CO<sub>2</sub>, including both fossil fuel and deforestation fluxes. Over the historical period (1750-2000), the GHG concentrations (except CO<sub>2</sub>) (Houghton et al., 2001) and carbon emissions (Marland et al., 2003; Houghton, 2003) follow the historical records. From 2000 to 2100, we use the SRES A2 scenario (Houghton et al., 2001) for both carbon emissions and GHG concentrations. After 2100, concentrations of all GHGs (except CO<sub>2</sub>) are kept fixed to their 2100 values, while CO<sub>2</sub> emissions from land use are set to zero and fossil fuel emissions decrease according to a bell-shaped curve so that they reach zero a few decades after 2200 (Figure 4). In parallel, each experiment is accompanied by a control simulation in which all the forcings are maintained at their 1750 values with no anthropogenic CO<sub>2</sub> emission. The climate-parameter set 11 used here corresponds to the parameter values used in previous studies.

The three carbon cycle parameter sets lead to contrasted responses of the atmospheric  $CO_2$  to the identical forcing (Figure 5, left). Maximal values of the atmospheric  $CO_2$  concentration differ by up to 169 ppmv between carbon sets 1 and 3 (Table II). By year 2500, they still differ by 133 ppmv, i.e. a relative difference of about 11%. With carbon cycle parameter sets 1 and 2, the land  $CO_2$  uptake outpaces the ocean uptake (Figure 5, middle), while the reverse happens with carbon parameter set 3.

The parameters related to the continental vegetation processes explain up to 87% of the difference in atmospheric CO<sub>2</sub> response between the various experiments. On such time scales, changes in the rain ratio or in the export production have a much smaller impact on the atmospheric CO<sub>2</sub>. The contribution of the rain ratio to the

maximum value of the atmospheric  $CO_2$  range is about 10%, while changes in remineralization depth explain about three percents. Such small changes (a few ppmv) are within the variability produced by the model and cannot be ascertained yet. All together, the three parameter sets allow us to obtain a change in the carbon climate sensitivity (as defined in Frank et al., 2010) of the order of 7% (Figure 5, right).

The third digit in the experiment name refers to the carbon cycle parameter set with relatively low (1), medium (2) or high (3) changes in atmospheric  $CO_2$  in response to the same emission scenario.

# 3. Last millennium climate

The climate of the last millennium is simulated for each of the 27 parameter sets involving climatic and carbon cycle parameters. All the simulations start at year 500 from an equilibrium state at that time and are run in transient mode until year 2000. Two different methods are used for the evolution of the atmospheric CO<sub>2</sub> concentration, either diagnostic or prognostic. In the diagnostic mode (Conc), the atmospheric CO<sub>2</sub> concentrations are prescribed according to Antarctic records until 1750 (Fluckiger et al., 2002; Monnin et al., 2004; Siegenthaler et al., 2005; Meure et al., 2006), according to Enting et al. (1994) between 1750 and 1990, and according to GLOBALVIEW-CO2 (2006) after 1990 (Figure 6). For the prognostic mode (Efor), the atmospheric CO<sub>2</sub> concentration is computed by forcing the model with emissions of CO<sub>2</sub> from fossil fuel burning (Figure 7; Marland et al., 2003). Both simulations also take into account land use changes related with human activities as in Goosse et al. (2005) (percentage of crops; Ramankutty and Foley, 1999). The scenario of historical changes in global land cover developed by Ramankutty and Foley (1999) starts only in 1700. We hypothesize that the land cover changes evolved linearly from its natural state in 500 to the estimated values in 1700. Moreover, we assume that croplands replace only forests, as long as there is a forest fraction. Furthermore, desert and forest (except for the part replaced by crops) keep their original extent at year 500. This scenario was used in a model intercomparison exercise aiming at analysing the response of six EMICs, including ECBilt-CLIO-VECODE, to historical deforestation (Brovkin et al., 2006).

In addition to the atmospheric  $CO_2$  concentration, either prescribed or computed by the model, the transient simulations are forced by the volcanic activity (Crowley,

2000), the solar activity (Muscheler et al., 2007), the Earth orbital parameter changes (Berger, 1978; Bretagnon, 1982) and changes in concentrations of GHGs other than  $CO_2$  (Prather et al., 2001; Houghton et al., 1990 and updates). The effect of sulphate aerosols is accounted for through a modification in surface albedo, as suggested by Charlson et al. (1991) (scenario S1).

The simulations reveal that all our parameter sets lead to a reasonable climate on millennial time scale. Although none of the simulations is perfect, none of them is in complete disagreement with available climate observations (or reconstructions). As illustrated by the simulated NH surface temperature in Conc and the atmospheric CO<sub>2</sub> concentration simulated in Efor (Figure 6), all simulations show relatively good agreement with reconstructions. Furthermore, this figure shows a strong consistency between all the simulations (and all the parameter sets) over the whole duration of the simulations. The comparison between simulated and reconstructed global variables representative of the climate evolution over the last millennium will not allow us to rank the ability of the parameter sets to properly simulate the climate reconstructions over that time scale are large, and the variability of the model may prevent an accurate comparison. Therefore, we decided to focus on the most recent part of these simulations (i.e. the last century; see main paper) for which some accurate measurements of climatic variables are available.

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Climatic parameter set	λ2 (m)	λ4 (m)	amplw	explw	albocef	albice	avkb	CorA
112	0.125	0.070	1.00	0.3333	1.000	0	1.0	-0.0850
122	0.120	0.067	1.00	0.4	0.900	0	2.0	0.0000
212	0.125	0.070	1.00	0.4	0.900	0	1.5	-0.0850
222	0.125	0.070	1.00	0.4	0.900	0	1.5	-0.0425
312	0.131	0.071	1.00	0.5	0.950	0	2.5	-0,0850
322	0.125	0.070	1.05	0.5	0.900	0	1.5	-0.0425
412	0.131	0.071	1.10	0.5	0.900	0	2.5	-0.0850
512	0.131	0.071	1.30	0.5	1.050	0.02	2.0	-0.0850
522	0.125	0.070	1.30	0.5	1.000	0.02	1.5	-0.0425

Table I: The nine 'climatic' parameter sets. The experiment names (column 1) are explained in Table I in the main paper. Parameters  $\lambda 2$  and  $\lambda 4$  (columns 2 and 3) are applied in the Rayleigh damping term of the equation of the quasi-geostrophic potential vorticity. The coefficients amplw and explw (columns 4 and 5) are used in the longwave radiative scheme to compute anomaly in humidity (see text). The uncertainties in the albedo of the ocean and sea ice are accounted for through albcoef (column 6) and albice (column 7). The minimum vertical diffusion coefficient in the ocean is scaled according to avkb (column 8). CorA is a correction factor for the distribution of precipitation over the ocean (column 9).

Carbon	βg	βt	$lpha_{ ext{diatom}}$	$\alpha_{\text{others}}$	$\Psi_{\text{zoo}}$	Atm.	CO <sub>2</sub>
parameter						(ppmv)	
set						Max	2500
							A.D
1	0.14	0.50	-0.750	-0.950	0.10	1146	877
2	0.36	0.36	-0.858	-0.858	0.22	1202	918
3	0.14	0.22	-0.648	-0.648	0.22	1315	1010

Table II: Model parameter sets for the carbon cycle and their effect on the CO<sub>2</sub> response. These parameters influence the continental vegetation fertilization effect ( $\beta$ g and  $\beta$ t; columns 2 and 3), the vertical flux of POM ( $\alpha_{diatom}$  and  $\alpha_{others}$ , columns 4 and 5), and the buildup of calcium carbonate shells ( $\Psi_{zoo}$ , column 6). Columns 7 and 8 give the maximum value of the annual mean atmospheric CO<sub>2</sub> concentration and its value at year 2500 from the transient simulations (see text) with the three carbon cycle parameter sets. Climatic parameter set 11 is used here.

Experiment				
name				
Exyz	Pre-industrial equilibrium:			
	No volcanic eruption, GHG as in 1750, TSI <sup>1</sup> = 1365 Wm <sup>-2</sup>			
Exyz2CO	Two times CO <sub>2</sub> scenario:			
	Starting from Exyz			
	Forcings as in Exyz except for the atmospheric CO <sub>2</sub>			
	concentration (Figure 2).			
ExyzHYS	Water hosing simulation:			
	Starting from Exyz			
	Forcings as in Exyz except for a freshwater perturbation applied			
	in the North Atlantic (Figure 3).			
ExyzTRA	Transient simulation from 1750 to 3000 starting from Exyz.			
	Forcings: orbital parameters, changes in concentration of GHGs			
	other than CO <sub>2</sub> , anthropogenic emissions of CO <sub>2</sub> (both fossil fuel			
	and deforestation fluxes).			

Table III: Summary of the major features of the different types of simulations discussed in the Supplement. See Table I in the main paper for the definition of xyz.

<sup>&</sup>lt;sup>1</sup> TSI = Total Solar Irradiance

					Equilibrium		
Name	TCR	CS	Ceff	Equi	MOC	MOC	Ts
	(1)	(2)	(3)	(4)	sensitivity (5)	(6)	(7)
	°C	°C	°C	°C	%	Sv	°C
112	0.7	1.6	1.6	1.6	-16	28.4	16.1
122	0.8	1.8	1.9	2.0	-52	17.3	15.8
212	0.8	2.0	2.2	2.2	-29	25.6	15.8
222	0.9	2.1	2.2	2.3	-57	21.5	15.6
312	0.9	2.5	2.7	2.9	-20	25.1	16.4
322	1.0	2.9	3.1	3.3	-54	20.9	15.7
412	1.1	3.2	3.5	3.7	-25	24.0	15.9
512	1.5	3.9	4.4	4.5	-34	23.9	16.1
522	1.5	4.1	4.8	4.8	-51	19.9	15.5

Table IV: Main features of the model climatic parameter sets with carbon cycle parameter set 2:

(1) increase in global annual mean surface temperature after 70 years in the doubling  $CO_2$  experiment from the equilibrium value;

(2) increase in global annual mean surface temperature after 1000 years in the doubling  $CO_2$  experiment from the equilibrium value;

(3) the effective climate sensitivity according to Gregory et al. (2002) (see also Goelzer et al., 2010);

(4) The equilibrium response in global annual mean surface temperature is computed after 2000 years for the parameter sets 112, 122, 212 and 222, and after 3300 years for the parameter sets 312, 322, 412, 512 and 522.

(5) percentage of decrease in the meridional overturning streamfunction after 1000 years in the water hosing experiment;

(6) strength of the meridional overturning streamfunction in the North Atlantic (Sv) at equilibrium in the pre-industrial experiment;

(7) annual mean global surface temperature (°C) at equilibrium in the pre-industrial experiment;

(8) minimum (min) and maximum (min) of sea ice extent in the Northern Hemisphere (10<sup>6</sup>km<sup>2</sup>) at equilibrium in the pre-industrial experiment;

(9) globally averaged, annual mean temperature of the ocean (°C) at equilibrium in the preindustrial experiment.

CO <sub>2</sub> data				
Series	References			
Taylor Dome	Indermühle et al., 1999			
Law Dome	Etheridge et al., 1998			
Siple	Neftel et al., 1994			
South Pole	Siegenthaler et al., 2005			
D47	Barnola et al., 1995			
D57	Barnola et al., 1995			
DML	Siegenthaler et al., 2005			
Temperature reconstructions				
Series	References			
B2000	Briffa, 2000; calibrated by Briffa et al., 2004			
BOS2001	Briffa et al., 2001			
DWJ2006	D'Arrigo et al., 2006			
ECS2002	Esper et al., 2002; recalibrated by Cook et al., 2004			
HCA2006	Hergel et al., 2006			
JBB1998	Jones et al., 1998; calibrated by Jones et al., 2001			
MBH1999	Mann et al., 1999			
MJ2003	Mann and Jones, 2003			
MSH2005	Moberg et al., 2005			
PS2004	Pollack and Smerdon, 2004; reference level adjusted following Moberg et al., 2005			
RMO2005	Rutherford et al., 2005			

Table V: References for the atmospheric CO<sub>2</sub> concentration (top) and temperature reconstructions (bottom) presented in Figure 6.



Figure 1: Zonally averaged surface temperature (°C), annual mean precipitation (cm/yr), and tree fraction (%) simulated for the pre-industrial time according to the nine climatic parameter sets. Observations are in black (Brohan et al. (2006) for temperature; Xie and Arkin (1996, 1997) for precipitation ; http://www.monsoondata.org:9090/dods/gswp/grid/fixed/classfrac\_igbp for tree fraction). Carbon cycle parameter set 2 is used here. The colour code for the climatic parameter sets is given in the figure. Carbon cycle parameter set 2 is used here.



Figure 2: Atmospheric  $CO_2$  concentration in the perturbation scenario (left) and time evolution of the global annual mean surface temperature response to this perturbation according to the selected model parameter sets (right). Temperature is presented as deviation from the initial value. The colour code for the climatic parameter sets is given in the figure. Carbon cycle parameter set 2 is used here.



Figure 3: Freshwater forcing in the North Atlantic in the perturbation scenario (left) and time evolution of the maximum of meridional overturning streamfunction below the Ekman layer in the Atlantic Ocean according to the selected model parameter sets in response to this perturbation (right). MOC is the absolute value. The colour code for the climatic parameter sets is given in the figure. Carbon cycle parameter set 2 is used here.



Figure 4: CO<sub>2</sub> emission scenario used to assess the sensitivity of the carbon cycle to the different carbon cycle parameter sets (see description of the scenario in the text). It includes both fossil fuel emission and fluxes related to land use change.



Figure 5: Evolution of the annual mean atmospheric  $CO_2$  concentration (ppmv) with time (left), terrestrial carbon inventory versus ocean carbon inventory (both in GtC) (middle) and atmospheric  $CO_2$  versus the global annual mean surface temperature (right) for the different carbon cycle parameter sets. The dashed line in the middle panel represents the 1:1 slope. Inventories are presented as anomalies with respect to the control run. The same color code is used in each panel, i.e. black for carbon cycle parameter set 1, green for set 2 and red for set 3. Climatic parameter set 11 is used here.



Figure 6: Evolution of the atmospheric  $CO_2$  concentration (ppmv) (left) and the Northern Hemisphere annual mean surface temperature (°C) over the last millennium (in year AD) as simulated by LOVECLIM according to the 27 parameter sets (brown). Results are displayed for Efor-simulations in the case of the atmospheric  $CO_2$  concentration and for Concsimulations in the case of temperature.  $CO_2$  concentration measured in Antarctic ice cores is shown for comparison. The full black line is the scenario of atmospheric  $CO_2$  concentration used in the Conc-simulations. Temperatures are expressed as anomalies from their 1500 to 1899 means. They are smoothed with a 31-yr window. Temperature reconstructions using multiple climate proxy records (see Jansen et al. (2007) for details) are in colour lines. The individual series are identified in Table V.



Figure 7: Evolution over the last centuries (in year AD) of the emission of  $CO_2$  (GtC/yr) from fossil fuel burning as prescribed in Efor simulations (Marland et al., 2003).