Model and experiments

The model we used to simulate the last nine interglacials is LOVECLIM\textsuperscript{10}, with the atmosphere (ECBilt), the ocean and sea-ice (CLIO) and the terrestrial biosphere (VECODE) components being interactively coupled. ECBilt is a quasi-geostrophic potential vorticity atmospheric model with 3 levels and a T21 horizontal resolution (Opsteegh et al., 1998). CLIO consists of an ocean general circulation model coupled to a comprehensive thermodynamic–dynamic sea-ice model (Goosse and Fichefet, 1999). Its horizontal resolution is 3° by 3°, and there are 20 levels in the ocean. VECODE is a reduced-form model of vegetation dynamics and of the terrestrial carbon cycle (Brovkin et al., 1997). It computes the evolution of the vegetation cover described as a fractional distribution of desert, tree and grassland at the same resolution as that of ECBilt. LOVECLIM has been used in a large number of studies for the past, present and future climate. It capably reproduces the major characteristics of the observed climate both for present-day conditions and for key periods of the past\textsuperscript{10}. In relation to some key features analyzed in this paper, the sea-ice extent/concentration and the maximum overturning streamfunction in the North Atlantic and Southern Ocean, as well as the location of deep convections simulated by LOVECLIM are in relatively good agreement with observations and with results of other models (Driesschaert, 2005; Goosse et al 2010\textsuperscript{10}). The spatial distribution pattern of deep-ocean temperature is well simulated by LOVECLIM, but it is generally cooler than the observation in particular in the Pacific. The capability of simulating large scale oceanic features has allowed Duplessy et al. (2007) to simulate successfully the deep ocean of the last interglacial. The large scale climate response of LOVECLIM compares favorably with that of other general circulation models using the same prescribed interglacial forcings (Muri et al., 2011; Herold et al., 2012; Nikolova et al., 2012).

The strategy for selecting insolation and greenhouse gases (GHG) forcings for each of the last nine interglacials is the same as in Yin and Berger\textsuperscript{11}. Firstly, the interglacials were identified by their peaks in the benthic δ\textsuperscript{18}O stack\textsuperscript{1}. Over the last 800 kyr, the interglacials are identified as Marine Isotopic Stages (MIS) 1 to 19. Secondly, following the hypothesis that an interglacial is associated with a maximum summer insolation in the Northern Hemisphere (NH), the astronomical parameters (Berger, 1978) were taken at the dates when NH summer occurs at perihelion just preceding the δ\textsuperscript{18}O peaks. This means that the longitude of the perihelion is fixed and the climatic precession is equivalent to eccentricity. These dates correspond quite well to the peaks of the δ\textsuperscript{18}O stack if we accept that the response time of the climate system to the astronomically-induced insolation is a few thousands of years. This lag between NH summer at perihelion and the peak of the interglacials also means that warm climates occur when NH autumn is close to perihelion. Thirdly, for the greenhouse gas forcing, CO\textsubscript{2}\textsuperscript{3}, CH\textsubscript{4} (Loulergue et al., 2008) and N\textsubscript{2}O (Schilt et al., 2010) were taken into account in our simulations, and their concentrations at the date of CO\textsubscript{2} maximum of each interglacial were taken. Table S1 gives the eccentricity, obliquity and CO\textsubscript{2} equivalent concentration for each interglacial. In the first set of nine simulations, the GHG was kept constant to the average of the nine interglacials in order to determine the pure impact of insolation. In the second set of nine simulations, which were used to compare simulated results with proxy data, both insolation and GHG vary. Each simulation was 5000-year long allowing the deep ocean and large scale ocean circulation to reach equilibrium, and the last 500-year average climatology was analyzed.

Model-proxy comparisons

Model-data comparison is often made in order to control both the simulated results and proxy-based reconstructions. This has been done for the simulated surface climatologies of the nine interglacials in Yin and Berger\textsuperscript{11} and Berger and Yin (2011). Here we focus on the oceanic data although the number of proxies for ocean circulation and deep-sea temperature is quite limited as compared to
the number of proxies for the surface climates. The results of the simulations where both insolation and GHG vary between the interglacials are used to compare with proxy data.

The impact of CO₂ on the oceanic features has been analyzed in many studies, and so is not discussed in detail here. In short, lower CO₂ concentration weakens the southern westerlies and consequently weakens the wind-driven upwelling. It causes stronger convection in the North Atlantic and in the Antarctic marginal seas. This intensifies the North Atlantic Deep Water (NADW) formation and the deep-water formation along the Antarctica shelf although the northward transport of the AABW near the bottom is reduced and associated with stronger southward transport of the NADW. The CO₂-induced changes simulated by our model are well in agreement with those summarized in the IPCC report for the future climate projections (Meehl et al., 2007). When both insolation and CO₂ are allowed to vary, the ventilation in the Southern Ocean and the AABW formation remain stronger in the pre-MBE interglacials. However, the northward export of AABW near the bottom is generally weaker due to reduced CO₂ concentration, except for MIS-15 when compared to MIS-1, 7 and 11 because of the dominant role of its insolation.

The main features of the last interglacial shown in Duplessy et al (2007), like the NADW warming and the bottom water cooling as compared to present-day are reproduced as well in our MIS-5 simulations. The AABW of MIS-5 is much stronger than at Pre-Industrial (PrI) (the maximum streamfunction in the vertical cell increases by 32% as compared to PrI), confirming the proxy-based reconstruction of Duplessy et al (1984). On the contrary, the MIS-5 NADW is reduced with its maximum streamfunction being 7% lower than PrI. As MIS-5 and PrI have a similar CO₂ concentration, their difference is mainly caused by the difference in their insolation. The much higher NH summer insolation during MIS-5 reduces significantly the NH sea-ice concentration and increases the temperature of the NADW source water all year round, leading to a weaker but warmer NADW. Our model also simulates less sea ice and warmer sea-surface temperature over the Southern Ocean during MIS-5 than PrI. As a consequence, the net evaporation is largely increased leading to saltier and finally to denser surface water. Moreover, the southern westerlies largely increased during MIS-5 as compared to PrI. Such denser surface water and stronger westerlies are responsible for the stronger AABW formation during MIS-5.

The difference in the NADW intensity between the interglacials is relatively small (the largest difference of about 3% in the maximum overturning streamfunction is between MIS-13 and MIS-9). Nevertheless, the overturning streamfunction, ventilation age and mixed layer depth in the North Atlantic consistently show that the NADW is stronger during MIS-13, 11, 17 and 7 than during the other interglacials. This is in good agreement with the NADW proxy of Raymo et al (1997) which shows that during the past 800 kyr, the greatest NADW occurs during MIS-13, 17 and 7. Stronger NADW during MIS-13 and MIS-11 is also indicated in the δ¹³C record from the North Atlantic (Flower et al., 2000). The strong NADW during MIS-13 was suggested to be a potential explanation for the strong East Asia summer monsoon recorded in the Chinese loess (Guo et al., 1998). Yin and Berger¹¹ show that the northern high latitudes temperature is more controlled by insolation than by CO₂ and more by obliquity than precession. The obliquity of MIS-13, 11, 7 and 17 is smaller than the obliquity of the other interglacials leading to cooler insolation-induced northern high latitudes (see Fig. 7a in Yin and Berger¹⁸) and finally to stronger NADW formation (Supplementary Fig. 1). The low CO₂ of MIS-13 and MIS-17 reinforces this cooling leading to these two interglacials being the coolest ones of the past 800 kyr. The higher CO₂ of MIS-7 and MIS-11 slightly counteracts this cooling but these two interglacials remain cooler than MIS-1, 5, 9, 15 and 19 in the northern high latitudes (see Fig. 7b in Yin and Berger¹¹). This northern high-latitude cooling of MIS-13, 11, 17 and 7 contributes significantly to increase the surface water density in the North Atlantic and finally to stronger NADW formation.

In the deep tropical Pacific, the relative magnitude of the simulated temperature of the nine
interglacials agrees with the estimated one of Siddall et al\textsuperscript{17} in the sense that the pre-MBE interglacials (except MIS-19) are cooler than the post-MBE ones, MIS-7 is the coolest one among the post-MBE and MIS-13 and MIS-17 are the coolest among all the nine interglacials. The deep-sea temperature estimated by Elderfield et al\textsuperscript{16} based on the Site 1123 located at east of New Zealand also shows cooler pre-MBE interglacials in agreement with my modeling results. However, although uncertainties exist in the proxy reconstruction, the difference in temperature between interglacials seems to be underestimated in our model which is probably due to the modelled deep Pacific being too cold to allow much variation between the interglacials. The deep-sea cooling during the pre-MBE interglacials results from a conjunction effect of the insolation-induced cooling (see main text) and the cooling caused by their lower CO\textsubscript{2} concentration.

A significant northward migration of the Antarctic Polar Front (APF) during the pre-MBE interglacials as compared to the post-MBE ones was suggested by Kemp et al\textsuperscript{9} based on the laminated diatom mat deposits in the South Atlantic, and it was further hypothesized to contribute to the lower CO\textsubscript{2} concentration during the pre-MBE interglacials. Here we use the surface density latitudinal gradient to trace the position of APF in our simulations. The largest difference in the APF position between the interglacials is located in the Atlantic sector (Supplementary Fig. 5a). It shows indeed that the APF (between 60°S and 48°S) shifts significantly northward during the coolest interglacials MIS-13 and MIS-17 and shifts southward during the warmest ones MIS-9 and MIS-5. The difference between MIS-1, 11 19, 7 and 15 is not obvious probably due to a limited spatial resolution of the model which does not allow to distinguish the smaller difference between these five interglacials. The difference in the APF location between the pre- and post-MBE interglacials is however not observed under the pure impact of insolation (Supplementary Fig. 5b). But it is clearly shown in an additional set of simulations where only the impact of CO\textsubscript{2} was taken into account (Supplementary Fig. 5c), with a higher CO\textsubscript{2} leading to a more south location of the APF. This is due to the fact that the position of APF is related to the sea-ice extent and sea-surface temperature in the Southern Ocean which is dominated more by CO\textsubscript{2} than by insolation\textsuperscript{11}. Lower CO\textsubscript{2} leads to larger sea-ice extent and cooler sea-surface temperature during the pre-MBE interglacials, a modeling result confirmed by proxies\textsuperscript{6,7,8}. This leads finally to a northward migration of the APF. Our modelled CO\textsubscript{2}-induced migration of the APF between the pre- and post-MBE interglacials implies that the southward migration of the APF during the post-MBE interglacials observed by Kemp et al\textsuperscript{9} might be more an effect of their higher CO\textsubscript{2} concentration instead of a cause as has been speculated, although the feedback mechanisms between the two makes the cause difficult to be separated from the effect.

An intensified deep western boundary current in the southwest Pacific during the post-MBE interglacials as compared to the pre-MBE ones is indicated by the proxy flow-speed data of Hall et al (2001). The model simulates correctly all the western boundary currents in the deep ocean with the southwest Pacific one being the strongest. In agreement with the result of Hall et al (2001), our model simulates a stronger western boundary current in the deep Pacific during the post-MBE interglacials. This is mainly due to their higher CO\textsubscript{2} concentration which strengthens the northward export of the AABW near the bottom. The order of the intensity of this current is therefore similar to that of the global annual mean surface temperature which is dominated by CO\textsubscript{2}\textsuperscript{11}, decreasing from MIS-9, 5, 11, 1, 19, 7, 15, 13 to MIS-17 with MIS-9 being 30% stronger than MIS-17.
Supplementary References


Supplementary table and figures:

Table 1  Eccentricity, obliquity (Berger, 1978) and CO₂ equivalent concentration of the last nine interglacials. 65°N summer solstice insolation is given as an indicator of insolation strength in the Northern Hemisphere.

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<th>MIS</th>
<th>eccentricity</th>
<th>obliquity</th>
<th>CO₂eq (ppmv)</th>
<th>65°N summer solstice Insolation (Wm⁻²)</th>
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**Figure 1.** Insolation-induced annual mean difference between each of the pre-MBE interglacials and each of the post-MBE ones in the meridional overturning streamfunction in the Southern Ocean and in the Atlantic. The separation of the two oceans is located at 32°S. Y-axis indicated depth in meters. The largest change occurs in the bottom water formation adjacent to Antarctica, with the maximum difference reaching 17% between MIS-15 and MIS-11.
**Figure 2.** Insolation-induced annual mean mixed layer depth and ventilation age in the Southern Ocean for each individual interglacial. The ventilation age is averaged vertically from 850m down to 5126m in the Southern Ocean from 60°S to 80°S, and from 850m down to 3661m in the North Atlantic, a method similar to that used in de Boer et al (2007). The two exceptions, MIS-7 and MIS-19, are indicated in white bars.
**Figure 3.** Insolation-induced zonal and annual mean differences between each of the pre-MBE interglacials and each of the post-MBE ones. They are for: a, surface density; b, salinity; c, sea-ice concentration; d, net evaporation.
Figure 4. Insolation-induced difference between each of the pre-MBE interglacials and each of the post-MBE ones in annual mean temperature of the global ocean. Y-axis indicates depth in meter.
**Figure 5.** Zonal and annual mean surface density latitudinal gradient in the South Atlantic from 50°W to 0°. They are for the simulations: **a**, where both insolation and CO₂ vary; **b**, where only insolation varies and CO₂ is kept constant; **c**, where CO₂ varies but insolation is kept constant.

**Figure 6.** Linear regressions between eccentricity, obliquity and the Equator-South Pole insolation gradient. They are: **a**, between eccentricity and the insolation gradient of a period from May to September; **b**, between obliquity and the insolation gradient of a period from October to April. The insolation is calculated using the total solar irradiation divided by the number of days during each season. The Fortran program made by Berger et al (2010) is used for calculating the total irradiation over a period of time using elliptic integrals.