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How cold was Europe at the Last Glacial Maximum? A synthesis of the progress achieved since the first PMIP model-data comparison

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Abstract

The Last Glacial Maximum has been one of the first foci of the Paleoclimate Modelling Intercomparison Project (PMIP). During its first phase, the results of 17 atmosphere general circulation models were compared to paleoclimate reconstructions. One of the largest discrepancies in the simulations was the systematic underestimation, by at least 10°C, of the winter cooling over Europe and the Mediterranean region observed in the pollen-based reconstructions. In this paper, we investigate the progress achieved to reduce this inconsistency through a large modelling effort and improved temperature reconstructions. We show that increased model spatial resolution does not significantly increase the simulated LGM winter cooling. Further, neither the inclusion of a vegetation cover compatible with the LGM climate, nor the interactions with the oceans simulated by the atmosphere-ocean general circulation models run in the second phase of PMIP result in a better agreement between models and data. Accounting for changes in interannual variability in the interpretation of the pollen data does not result in a reduction of the reconstructed cooling. The largest recent improvement in the model-data comparison has instead arisen from a new climate reconstruction based on inverse vegetation modelling, which explicitly accounts for the CO_2 decrease at LGM and which substantially reduces the LGM winter cooling reconstructed from pollen assemblages. As a result, the simulated and observed LGM winter cooling over Western Europe and the Mediterranean area are now in much better agreement.

1 Introduction

The aim of the first phase of the Paleoclimate Modelling Intercomparison Projet (hereafter PMIP1) was to assess the sensitivity of the atmosphere general circulation models (AGCMs) used to predict future climate change to very different conditions. The Last Glacial Maximum, which occurred 21 000 years ago, was chosen as a test for extremely cold conditions. Indeed this period is relatively well documented in terms of paleoclimate reconstructions to which model results can be compared. For instance, for Europe and western Siberia, models and reconstructions could be compared in terms of Mean Annual Temperature (MAT), Mean Temperature of the Coldest Month (MTCO), total annual precipitation (TAP) and moisture index (ratio of the mean annual actual evaporation to the mean annual potential evaporation). These bioclimatic parameters were computed from pollen assemblages using a statistical method based on a transfer function associating present Plant Functional Types to climatic data (Peyron et al., 1998; Tarasov et al., 1999). For instance, the analogues found for the Western Europe pollen assemblages in the original climatic reconstructions from Peyron et al. (1998) were located in tundra or very cold steppe environments, which resulted in very cold MTCO reconstructions.

To involve a large number of groups and models that were not always familiar with paleoclimates, the idea in PMIP1 was to follow an LGM experimental design as simple as possible (Joussaume and Taylor, 1995 and http://www-lsce.cea.fr/pmip/). The boundary conditions were set as follows: the CO_2 atmospheric concentration was fixed at 200 ppm from the Vostok measurements (Raynaud et al., 1993) and the orbital parameters were set to their 21 ky BP values (Berger, 1978); the ice-sheet elevation and extension, as well as the land-sea mask were prescribed using the ICE-4G reconstruction (Peltier, 1994); eight simulations used prescribed Sea Surface Temperature (SST) and sea-ice extension derived from CLIMAP (1981), while the eight others used a slab-ocean model to compute SST and sea-ice cover.

The comparison between these model results and pollen-based reconstructions over Europe and western Siberia showed a relatively good agreement for the MAT, but large discrepancies over western Europe and the Mediterranean area for MTCO and TAP (Kageyama et al., 2001). Since then, the data base used for the calibration of the Peyron et al. (1998) and Tarasov et al. (1999) transfer function method has been updated (Peyron et al., 2005) and new reconstructions for western Europe and the Mediterranean area are slightly warmer (Jost et al., 2005). However, the comparison of the PMIP1 model results with this updated MTCO reconstruction (Fig. 1) shows that the model-data disagreement remains, even if one takes into account the large error bars due to a poorly diversified vegetation (steppe-tundra) and a tolerance of this type of vegetation to large climatic amplitudes. This therefore raised a series of questions from both the models and the data points of view. In the present paper, we first review different model improvements: using a higher spatial resolution, investigating the role of missing interactions or feedbacks, such as from the biosphere and the ocean. We then investigate factors that were not included in the first pollen-based reconstructions: the potential impact of a different interannual variability and that of a much lower CO_2 at LGM, via the use of an Inverse Vegetation Model.

2 Increasing the spatial resolution of the models

Paleoclimatic reconstructions such as those based on pollens are relevant for a small area around the sites where they have been retrieved, compared to the usual resolution (i.e. a few hundred km) of the AGCM used in PMIP1. Therefore, the spatial resolution of the PMIP models can be considered as a weak point when comparing model results to data, particularly in areas of complex coastlines and topography such as southern Europe and around the Mediterranean Basin. For sites close to mountain ranges such as the Pyrenees and the Alps, local climate can be very different from the climate simulated in the corresponding grid box of the models. During the LGM, these mountain ranges were partly covered by large glaciers, which can affect the local atmospheric circulation but are not represented in the GCMs. Increasing the models' resolution should improve the representation of a given climate. However, it is unclear that the sensitivity of the models to changes in their boundary conditions will be sensitive to their resolution. This question has been investigated by comparing the simulations of three AGCMs at low and high resolution (Jost et al., 2005), the high resolution being achieved through three different methods. All simulations were run according to the PMIP1 protocol. CCSR1 provided a run with a global T106 resolution (low resolution: T21), whereas the HadRM regional model was nested within the HadAM global AGCM over a domain including the North Atlantic and Europe, and LMDZ used a stretched grid version, with higher resolution over Europe (low resolution: 72 points in longitude x 46 points in latitude, high resolution 144×108, the resolution reaching 60 km over Paris). All these models reach a spatial resolution of around 50 km. However in terms of MTCO (Fig. 2), there is not any convincing improvement for CCSR1 and LMDZ, while HadRM succeeds in simulating cooler winter temperatures than the low resolution model, those temperatures being in agreement with the reconstructions. However, the latter model simulates significant increases in precipitation over western Europe and the Mediterranean areas which are in total disagreement with the reconstructions. Therefore, Jost et al. (2005) argue that spatial resolution cannot explain the model – data MTCO discrepancy over Western Europe and the Mediterranean area.

3 Increasing the number of components of the climate system taken into account in the models

The use of AGCMs implied to provide the models with accurate surface conditions. For the oceanic conditions, as explained in the introduction, the CLIMAP (1981) data set was chosen. For the biosphere, to simplify the PMIP1 experimental design, the protocol specified not to impose any change except for the new continental grid points due to the LGM sea-level drop, which characteristics are obtained by averaging those of neighbouring land points. For other periods, taking more components than the sole atmosphere into account in the models has been necessary to improve the model – data comparison. For instance, for the simulation of the onset of the last glaciation 115 kyr ago, it has been shown that accounting for vegetation (e.g. de Noblet et al., 2006) largely amplifies the cooling related to the northern hemisphere decrease in summer insolation. Furthermore, including ocean interactions in the experimental design via the use of a coupled atmosphere-ocean GCM, Khodri et al. (2001) demonstrated that atmosphere and ocean feedbacks on the water cycle favoured the accumulation of snow in the northern high latitudes. For the Mid Holocene the extension of African monsoon to the North also needs to be amplified by ocean and biosphere feedbacks to be in better agreement with data (Braconnot et al., 2000).

3.1 Investigating the role of the biosphere

The PMIP1 LGM experiments all used the present vegetation cover. South of the Fennoscandian ice-sheet, the albedo could therefore be underestimated at locations where forest is present today. In European regions presently dominated by agriculture, the difference between prescribing the present vegetation and the LGM one (steppe, tundra) might not be very large in terms of climate, since these surfaces are not very different in terms of albedo or roughness length. Several studies have investigated the impact of an LGM vegetation on a simulated LGM climate (Crowley and Baum, 1997; Kubatzki and Claussen, 1998; Levis et al., 1999; Wyputta and McAvaney, 2001; Crucifix and Hewitt, 2005). This impact has been evaluated either with respect to present climate simulations using the actual or pre-industrial vegetation or to simulations using present potential vegetation. While the impact of prescribing or interactively computing LGM vegetation can be large in regions like Siberia, none of these studies shows a large impact (more than a few °C) on European temperatures. However, results are not often given in terms of MTCO, but rather, most often, in terms of Mean Annual Temperatures.

Here, to investigate the possible impact of using a present vegetation in an LGM climate simulation, we have computed the LGM vegetation cover associated with the high resolution LGM climate obtained with LMDZ as described in the previous section, using the ORCHIDEE Dynamic Global Vegetation Model (Krinner et al., 2005). This DGVM describes vegetation in terms of natural vegetation (competition between 10 Plant Functional Types) and agriculture (2 vegetation types) + bare soil. All vegetation types can co-exist in each model grid box and the vegetation is described through the fraction occupied by each vegetation type. The obtained vegetation for the LGM over western Europe is characterised by a poor vegetation cover, dominated by grasses rather than forests (Kageyama et al., 2005). We have then imposed the computed

LGM vegetation back to the LMDZ model, still at high resolution over Europe. Results in terms of temperature anomalies for MTCO are shown on Fig. 3. The impact of vegetation superimposes a cooling of 1 to 2°C for MTCO in western Europe. This does not by itself compensate for the model/data discrepancy but does act to slightly improve the model results. The region of simulated cooling is limited, showing that a vegetation compatible with the LGM climate does not act to systematically cool winter temperatures for the Last Glacial Maximum climate. For instance, including an LGM vegetation induces a warming between the Black and Caspian Seas (Fig. 3).

3.2 Investigating the role of the ocean

Several studies have shown that the CLIMAP SST and sea-ice cover data set was to be improved and new estimates have recently been made available (e.g. through the MARGO project, Kucera et al., 2005). On the other hand, models using different SST and sea-ice cover estimates for the LGM also helped in the quantification of the impact of possible errors in CLIMAP (Pinot et al., 1999). However, to avoid the problems related to prescribing SSTs at the global scale, a major effort has been achieved by several groups in the recent years to produce coupled atmosphere-ocean GCM simulations of the LGM climate. After some pioneering simulations (e.g. Hewitt et al., 2001; Kitoh et al., 2001; Peltier and Solheim, 2004), such experiments were recently run within the PMIP2 exercise (Paleoclimate Modelling Intercomparison Project, 2nd phase, Harrison et al., 2002; Braconnot et al., 2006). For this exercise, atmosphere-ocean GCM (AOGCM) simulations have been performed using 1) the same 21 ky BP insolation (Berger, 1978) 2) a decrease in the atmospheric CO_2 concentration reevaluated to 185 ppm (Monnin et al., 2001) and 3) the new global ice sheet reconstruction (ICE-5G) from Peltier (2004). Despite a rather large spectrum of ocean dynamics responses (Weber et al., 2006), including different sea ice extents in North Atlantic (for more detail, see Kageyama et al., 2006), Fig. 4 shows that the discrepancy between model and data over western Europe and Mediterranean Basin is unchanged. This result demonstrates the rather weak sensitivity of the simulated MTCO to large differences in LGM SST produced by the models involved in PMIP2, which is quite unexpected.

4 Investigating the impact of a different climate variability on vegetation

Usually, most PMIP model results have been analysed in terms of changes in the mean climate, defined as the average of climatic variables such as MTCO on around 15 years of simulation (for PMIP1, longer for PMIP2). Results from proxy-based reconstructions for the LGM have also been interpreted as changes in the mean climate as these proxies are calibrated against the modern climatology. The assumption has therefore been that interannual variability does not vary. However, Kageyama et al. (2006) show that in fact, the interannual variability in MTCO changes in the LGM simulations, compared to the CTRL ones. Most PMIP2 models simulate an increase in the amplitude of interannual MTCO variability, suggesting the occurrence of much cooler extreme episodes than if we assume a constant interannual variability. This could have an impact on the vegetation which is not taken into account in the present pollen data interpretation.

We have therefore performed preliminary sensitivity experiments to investigate under which conditions an agreement with the pollen data can be obtained. In particular, we test whether changes in interannual variability could lead to significant changes in vegetation, in comparison to changes in the mean climate. We use the Lund-Potsdam-Jena Dynamic Global Vegetation Model (LPJ DGVM, Sitch et al., 2003), run offline under different CO_2 and climatic forcings. For these preliminary experiments, the climate inputs (averages and variability) were computed from the IPSL_CM4 model PMIP2 simulations and the Climatic Research Unit (CRU) observations (Mitchell et al., 2004). The simulations are run for 5 sites in Western Europe and the Mediterranean Basin and the results in terms of leaf area index (LAI) are presented on Fig. 5.

Under present climate and CO_2 forcings (experiment 1), the model mostly simulates forests and very little steppe or tundra. We then search for conditions for which we obtain the steppe-tundra observed during the Last Glacial Maximum. We first use an

LGM CO_2 concentration of 200 ppm and keep the modern climatology to force LPJ (experiment 2). This results in a LAI reduction but does not significantly affect vegetation composition. In the following experiments, the CO_2 concentration is kept to its LGM value. The third scenario is a test to the change in interannual variability: the mean climatology is the modern one, but the temperature variability is increased by a factor of 3. For three of the five sites, this results in a change of the vegetation composition, but never in a dominant steppe or tundra. Steppe-tundra vegetation occur simultaneously in all sites in the last experiment, in which mean temperature anomalies 1.25 colder than the ones simulated by the IPSL_CM4 model and total annual precipitation anomalies 3 times smaller then simulated by IPSL_CM4 are included in the climatic forcing. In practise, in terms of temperature, this means that the temperature should be 5 to 10°C colder than simulated by the IPSL_CM4 model in order to be consistent with a steppe-tundra vegetation. This is only a necessary condition, since total precipitation anomalies also need to be 100 to 600 mm/year less than simulated by IPSL_CM4. Without setting such cooler and drier conditions in average, variability changes alone cannot induce steppe or tundra vegetation to be simulated. Similarly, when these mean conditions are set, changing variability does not affect simulated vegetation composition.

5 New temperature reconstructions using an inverse vegetation model

The statistical reconstruction methods previously used (Peyron et al., 1998; Tarasov et al., 1999; Jost et al., 2005) are built upon the assumption that plant-climate interactions remain the same through time, and the fact that the calibration is done on modern data implicitly assume that these interactions are independent of changes in atmospheric CO_2. This assumption may lead to a considerable bias, as polar ice core records show that the atmospheric CO_2 concentration was much lower in the LGM than in the present time (EPICA community members, 2004). Less carbon was then available for photosynthesis processes, and the effect of the reduced CO_2 levels is interpreted by statistical reconstructions methods directly as change in one or more climatic parameters. At the same time, a number of palaeoecological studies (Jolly and Haxeltine, 1997; Street-Perrott et al., 1997; Cowling and Sykes, 1999) have shown that plant-climate interactions are sensitive to atmospheric CO_2 concentration in the past, and that part of the observed changes may be attributed to changes in CO_2 levels alone. The only solution to solve this problem is to use a vegetation model in an inverse mode, i.e. to calculate some model inputs (climate) when model outputs are constrained by the pollen data and the CO_2 concentration is set to its correct value. This model inversion, which uses a Monte-Carlo-Markov-Chain algorithm to explore possible combinations of climate parameters, was applied to a few sites of Southern Europe by Guiot et al. (2000). The exploration of a large number of climate scenarios allows an assessment of the probability of different anomalies, and therefore the investigation of different possible combinations of climate parameters, that may result in similar vegetation. They showed that, whilst there remains a high probability of the same reduction in temperatures reconstructed by the statistical method (Peyron et al., 1998), there is an equal probability of an alternative warmer climate when the CO_2 concentration is set to 200 ppmv. It was, however, not possible to chose between these two scenarios.

Recently Wu et al. (2007) have improved the method by better constraining the model output with pollen data and in using a more recent version of the model (BIOME4, Kaplan et al., 2003). They applied this to the data used by Peyron et al. (1998) and Tarasov et al. (1999). Their results confirmed that several solutions were possible for the LGM climate in Western Europe where a mixture of steppes and tundra existed. As these biomes have no clear analogues today, reconstructions based on statistical methods will tend to choose the least poor match, or fail to find a match. In the dataset used by Peyron et al. (1998), these analogues were located in tundra or very cold steppes, resulting in very low reconstructed temperatures. In the improved dataset of Jost et al. (2005), the analogues selected were intermediate analogues in warmer steppes. In the inverse modelling results, Wu et al. (2007) showed that a significantly warmer climate was the most probable and that statistical method overestimated the MTCO anomalies by about 10°C, referring to a climate that has no modern analogue. Another interesting result is that the uncertainties were also underestimated with statistical methods (Fig. 6), again due to lack of well-constrained modern analogues for the LGM vegetation. The new reconstructions are now much closer to the simulated temperatures, and most of the simulations are within the error bars of the data. However, they remain too warm (Fig. 7) at least for MTCO, although the consistency is much better for MTWA and MAT.

6 Conclusion and perspectives

Thanks to the improved methodology based on inverse vegetation modelling (Wu et al., 2007), new estimates of the cooling over Western Europe at LGM are available. These estimates correspond to a reduced cooling compared to previous reconstructions, mainly because of the progress achieved in accounting for the low CO_2 impact. These new reconstructions are also characterised by an increase in the associated error bars, in comparison to the previous ones, in better agreement with ecophysiology. Figure 7 shows both the most recent reconstructions and the new simulations from PMIP2. On this plot, most models are now in the upper part of the error bars for MTCO. The preliminary results presented in Sect. 4 show that to be consistent with the observed LGM vegetation over Europe and the Mediterranean Basin, the IPSL_CM4 model would need to be cooler by 5 to 10°C. From Fig. 7, we can see that this indeed would result in a better agreement with the new MTCO estimates for this model. The results obtained via the two vegetation models are therefore consistent. This will need to be confirmed by extending the analyses presented in Sect. 4 to other models.

As we show in Sect. 3.1, the impact of vegetation is a 1 to 2°C cooling in MTCO for Western Europe. This could account for some of the remaining difference between the simulations and the reconstructions. Therefore, we can expect that next AOV GCM simulations to yield a better agreement with the new reconstructions.

In this work, we have focused on one variable for which the model-data discrepancies was very large: the temperature of the coldest month. This work should be followed by the analysis of another large model-data discrepancy: the representation of the total annual precipitation. Indeed, the first PMIP model-data comparison showed an important underestimation of the drying of western Europe and the Mediterranean Basin. This result will be re-evaluated in the light of new model simulations and new reconstructions as has been proposed here for the temperature of the coldest month. The next steps in modelling the Last Glacial Maximum climate will be to use fully coupled atmosphere-ocean-vegetation models, and then to include representations of the marine and terrestrial carbon cycles. This should lead to a better understanding of the reconstructions.

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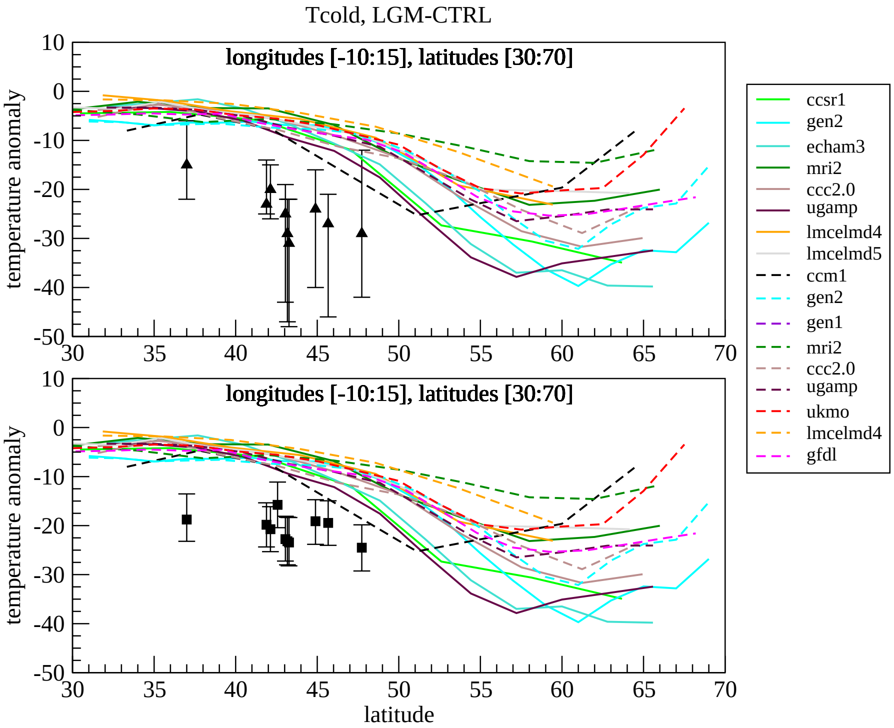

Fig. 1. **(a)** MTCO anomalies between the LGM and CTRL results for the PMIP1 models, compared to the MTCO anomalies reconstructed from pollen as described in Peyron et al. (1998). **(b)** Same MTCO anomalies but the data set has been updated as described in Jost et al. (2005).

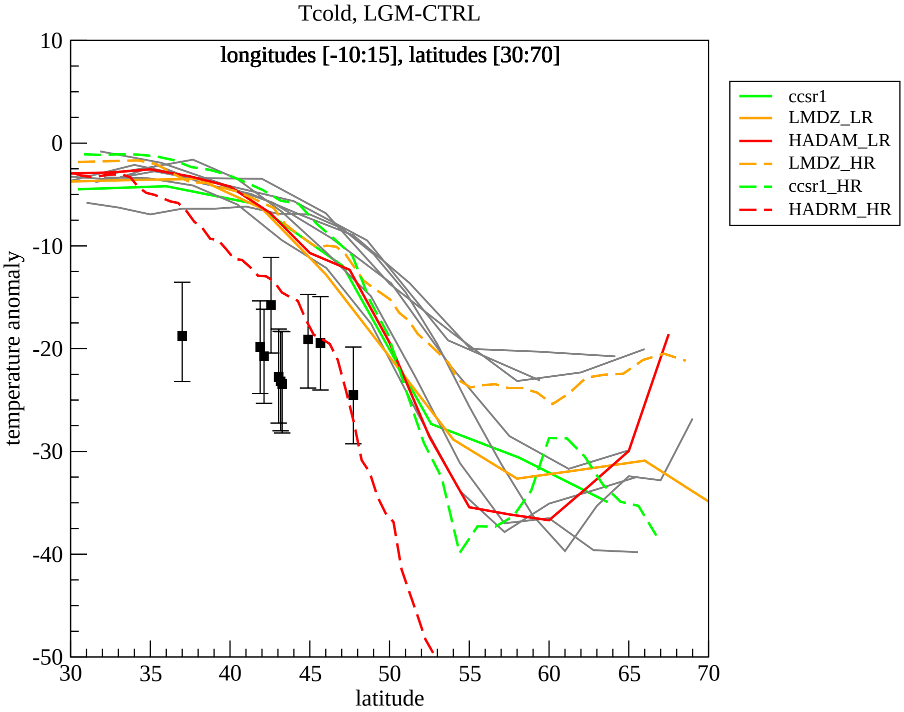

Fig. 2. MTCO LGM – CTRL anomalies for high resolution models, compared to their low resolution counterparts. The results from the PMIP1 prescribed SST experiments are shown in grey for comparison.

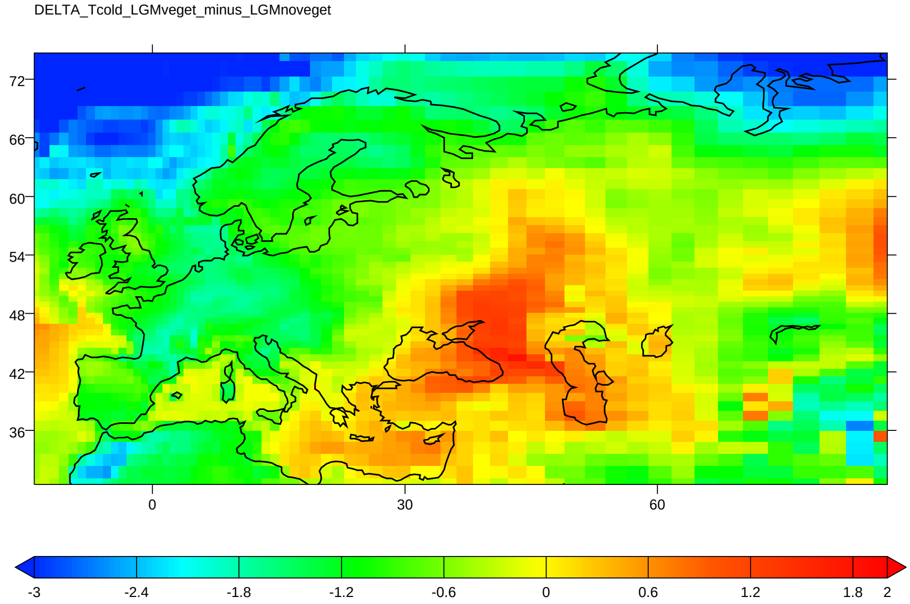

Fig. 3. MTCO anomalies between an LGM simulation run with PMIP boundary conditions (no change in vegetation) and an LGM simulation where the vegetation is computed by asynchronous coupling with the Dynamical Global Vegetation Model ORCHIDEE. Both these simulations have been performed with the LMDZ model with a stretched grid over Europe, in the version of Jost et al. (2005).

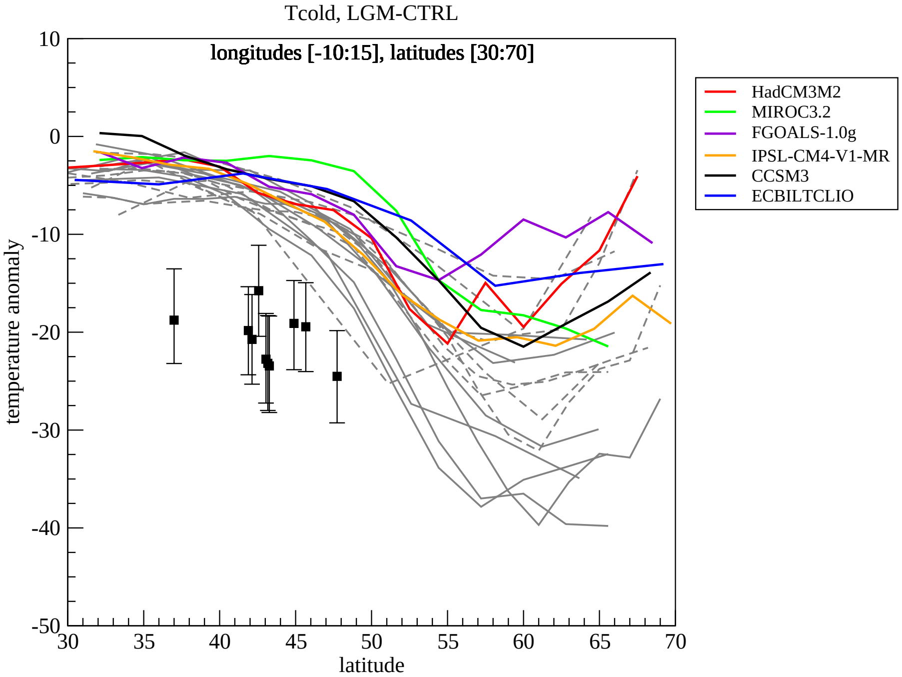

Fig. 4. MTCO anomalies between LGM and CTRL for the coupled OA GCM involved in PMIP2, compared to the Jost et al. (2005) pollen-based reconstructions. The PMIP1 model results from Fig. 1 are shown in grey for easier comparison.

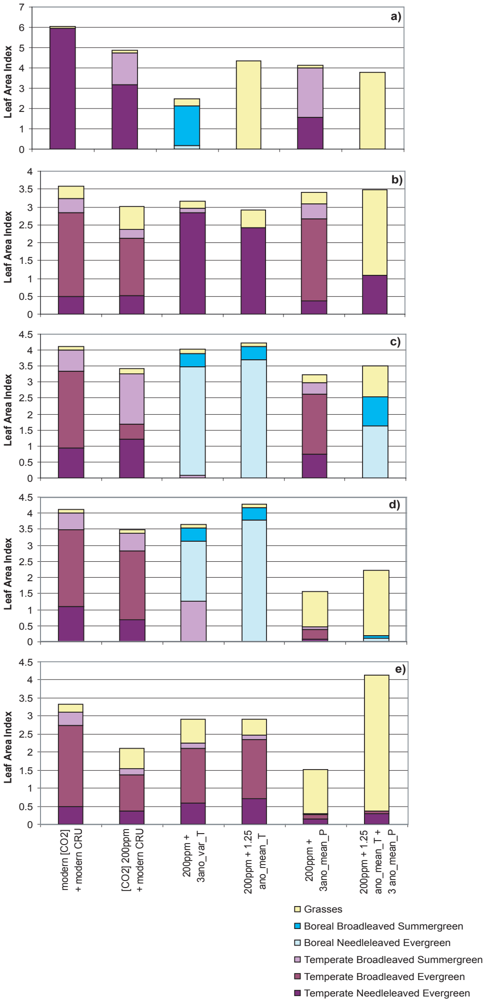
a)
b)
c)
d)
e)
Leaf Area Index
0
1
2
3
4
5
6
7
0
0.5
1
1.5
2
2.5
3
3.5
4
4.5
modern [CO2] + modern CRU
[CO2] 200ppm + modern CRU
200ppm + 3ano_var_T
200ppm + 1.25 ano_mean_T
200ppm + 3ano_mean_P
200ppm + 1.25 ano_mean_T + 3 ano_mean_P
Grasses
Boreal Broadleaved Summergreen
Boreal Needleleaved Evergreen
Temperate Broadleaved Summergreen
Temperate Broadleaved Evergreen
Temperate Needleleaved Evergreen

Fig. 5.

Fig. 5. Vegetation composition simulated for five western European sites and six different scenarios. Each graph corresponds to a site: **(a)** La Grande Pile in northern France, **(b)** Padul in Spain, **(c)** Monticchio in Italy, **(d)** Ioannina in Greece, and **(e)** Ghab in Syria. For each site, the following forcings have been applied to the LPJ Dynamic Vegetation Model: 1) modern climatology (i.e. modern climatological average and variability as described by the CRU dataset) and CO_2 (345 ppm), 2) LGM CO_2 (200 ppm), modern climatology (average and variability), 3) LGM CO_2, modern climatological average, temperature interannual variability increased by a factor of 3, 4) LGM CO_2, mean temperature average obtained by subtracting the mean temperature anomaly simulated by the IPSL model from the CRU climatological average, modern temperature variability, no other changes to the climatic forcing 5) LGM CO_2, total annual precipitation obtained by subtracting the anomaly simulated by the IPSL model from the CRU climatological average, no other change applied to the climatic forcing, 6) LGM CO_2, mean average temperature obtained by subtracting 1.25 times the LGM simulated anomaly to the CRU average, total precipitation obtained by subtracting three times the LGM simulated anomaly to the CRU average, no other change applied to the climatic forcing.

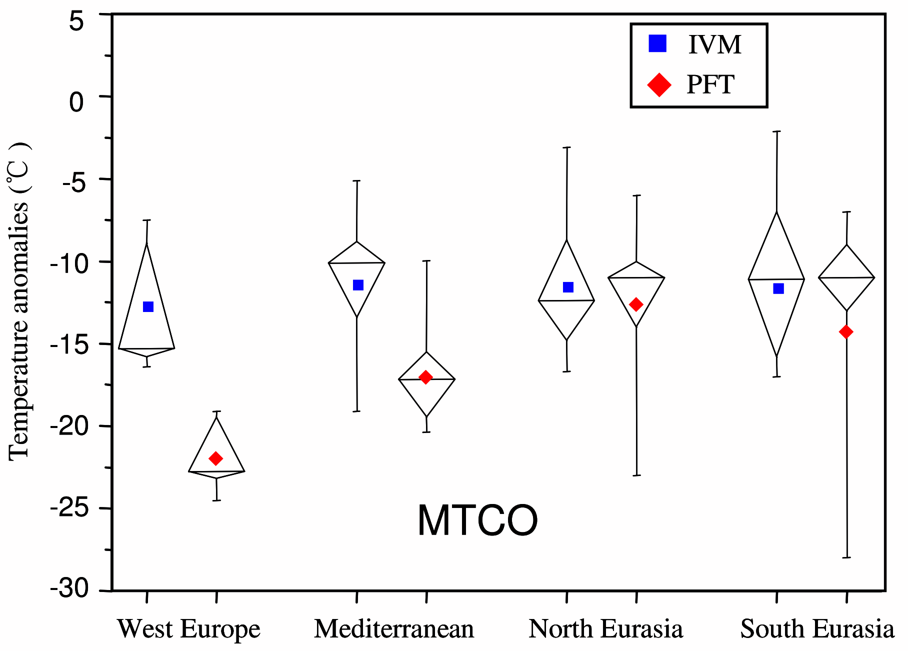

Fig. 6. Comparison of the MTCO reconstructions using the Plant Functional Type (Jost et al., 2005) and the Inverse Vegetation Modelling methods.

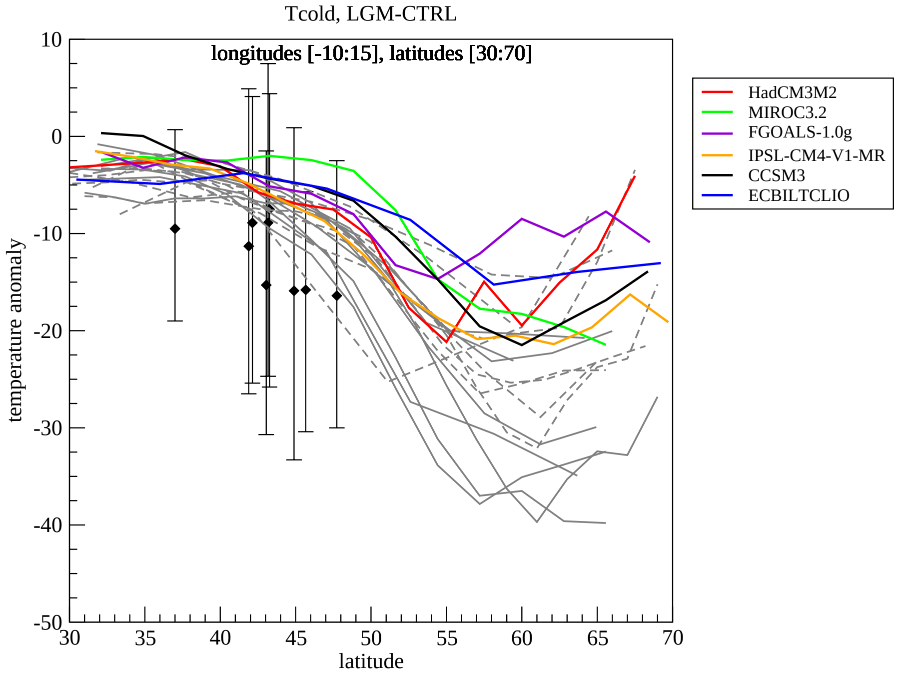

Fig. 7. MTCO anomalies as computed by PMIP1 and PMIP2 simulations compared to the new pollen-based reconstructions, using Inverse Vegetation Modelling.