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Antarctic temperature changes during the last millennium: evaluation of simulations and reconstructions

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\textbf{Abstract}

Temperature changes in Antarctica over the last millennium are investigated using proxy records, a set of simulations driven by natural and anthropogenic forcings and one simulation with data assimilation. Over Antarctica, a long term cooling trend in annual mean is simulated during the period 1000–1850. The main contributor to this cooling trend is the volcanic forcing, astronomical forcing playing a dominant role at seasonal timescale. Since 1850, all the models produce an Antarctic warming in response to the increase in greenhouse gas concentrations. We present a composite of Antarctic temperature, calculated by averaging seven temperature records derived from isotope measurements in ice cores. This simple approach is supported by the coherency displayed between model results at these data grid points and Antarctic mean temperature. The composite shows a weak multi-centennial cooling trend during the pre-industrial period and a warming after 1850 that is broadly consistent with model results. In both data and simulations, large regional variations are superimposed on this common signal, at decadal to centennial timescales. The model results appear spatially more consistent than ice core records. We conclude that more records are needed to resolve the complex spatial distribution of Antarctic temperature variations during the last millennium.

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1. Introduction

In Antarctica, the recent temperature trends are more spatially heterogeneous compared with other continental regions (Turner et al., 2005; Schneider et al., 2006; Hegerl et al., 2007; Chapman and Walsh, 2007; Monaghan et al., 2008a; Steig et al., 2009; ACCE, 2009; O’Donnell et al., 2011). Except over the Antarctic Peninsula where a large warming has been observed over the last 50 years (Vaughan et al., 2003; Turner et al., 2005), the few instrumental Antarctic air temperature records display an increase between 1960 and 1980 (AD, as all the dates given here) but a stabilization or a cooling over the following decades (Turner et al., 2005; Schneider et al., 2006; Chapman and Walsh, 2007; Monaghan et al., 2008a; Steig et al., 2009). For the period 1957—2006, the spatial reconstruction of Steig et al. (2009) exhibits a weak trend in East Antarctica and a strong temperature increase in West Antarctica, which has warmed by more than 0.1 °C per decade over this period. This warming in West Antarctica is particularly strong in spring (Schneider et al., 2012) and has been related to a forcing of the high latitudes by tropical sea surface temperature anomalies through atmospheric teleconnections (Ding et al., 2011, 2012; Schneider et al., in press). The increase in the Southern Annular mode index, partly due to stratospheric ozone and greenhouse gas forcings, also influences Antarctic
2. Overview of the climate changes over the past millennium in the southern hemisphere

At the scale of the southern hemisphere, the few available temperature reconstructions (Mann and Jones, 2003; Mann et al., 2008) suggest a cooling between the first part of the past millennium and the period 1500–1800 and a warming over the past 150 years. This might be interpreted as the southern hemisphere equivalent of the transition between the so-called Medieval Climate Anomaly and the Little Ice Age extensively analyzed for the northern hemisphere (e.g., Hughes and Diaz, 1994; Mann et al., 2008, 2009; Esper et al., 2009; Kaufman et al., 2009; Frank et al., 2010; Ljungqvist, 2010; Diaz et al., 2011). The timing is, however, slightly different in the southern hemisphere. Prior to the 20th century, the warmest conditions reconstructed by Mann et al. (2008) appear around 1300 in the southern hemisphere, compared to 950–1250 in the northern hemisphere. In South America, warm local summers are depicted from 900 to the middle of the 14th century followed by a sharp cooling (Neukom et al., 2011). The similarity with the reconstruction of Mann et al. (2008) may, however, be due to the use of common source proxy records. Warm summer periods are also reconstructed in South America between 1710 and 1820 and after 1940 (Neukom et al., 2011). A recent reconstruction for summer temperature in Australasia (Gergis et al., submitted for publication) agrees broadly with the general picture described above for the whole southern hemisphere and South America. By contrast, regional temperature reconstructions from Australia, Tasmania and New Zealand (e.g., Cook et al., 2002; Cook et al., 2006) do not generally display any clear transition from an equivalent of the MCA to the LIA.

Most Antarctic and subantarctic temperature related proxies display pronounced multi-decadal variations without a clear dominant common signal. A few oceanic cores close to the coast (e.g., Khim and Yoon, 2002 and Shevenell et al., 2011 close to the Antarctic Peninsula) suggest generally milder conditions during the first half of the past millennium than over the period 1500–1800. The number and the resolution of the available sediment cores in the Southern Ocean (e.g., Nielsen et al., 2004; Crosta et al., 2007; Anderson et al., 2009; Denis et al., 2010) is, however, too low to draw unequivocal conclusions at large-scale from oceanic data. The lake records along East Antarctica (see the review by Verleyen et al., 2011) and the majority of the Antarctic ice core records (e.g., Steig et al., 1998; Masson et al., 2000; Oerter et al., 2000; Mulvaney et al., 2002; Masson-Delmotte et al., 2004) do not show any significant long term temperature trend over the period 1000–1850. At these multi-centennial scales, marked regional differences have been reported. Lake sediment cores from the Northern Ocean together with those from the Ross Sea sector during the 12th and 13th centuries (Berthier et al., 2011). The same authors suggest that during the period 1500–1800, East Antarctica experienced cooler and drier conditions with higher wind speeds, cooler SSTs, more extensive sea-ice, and they inferred an increase in bottom water formation. Stenni et al. (2002), comparing an isotopic record obtained from a firn core drilled at Talos Dome with other East Antarctic ice core records, also suggest a cooler climate during this period, although not temporally synchronous between the different sites.

Borehole temperature measurements provide very useful information on past surface temperature trends over the last decades to centuries. In the Antarctic Peninsula area, they confirm the strong warming trend since 1900 depicted by instrumental, ice core and lake records (Barrett et al., 2009; Thomas et al., 2009; Zagorodnov et al., 2011; Sterken et al., 2012). On the West Antarctic Ice Sheet (WAIS) Divide, reconstruction from borehole temperature data show a cooling from the beginning of the past millennium until the 17th century, followed by a warming that intensifies during the 20th century (Orsi et al., 2012), the magnitude of the recent warming being in agreement with the estimates of Steig et al. (2009). The signals reconstructed from shallow borehole data appear less homogenous in the interior of Dronning Maud Land, East Antarctica, where high elevation sites display a warming of more than 1 °C over the last 50 years while no trend is detected at lower altitudes (Muto et al., 2011).

Similarly to the teleconnections observed during the instrumental period (see the introduction), high southern latitude temperature changes during the past millennium are expected to be closely related to modifications in large-scale atmospheric circulation patterns. The variability of ENSO has been extensively investigated but large uncertainties remain regarding its changes over the course of the past millennium (e.g., Mann et al., 2005; McGregor et al., 2010). At mid-latitudes, an intensification and/or a northward shift of the westerlies has been suggested after the 15th century in the south-east Pacific, over southern South America and over South Africa (Mobtadi et al., 2007; Moy et al., 2008; Moreno et al., 2009; Sepulvada et al., 2009; Lamy et al., 2010; Stager et al., 2012). However, the timing of the shift is not entirely consistent (within one century) amongst the various records, possibly because of the superimposition of large multi-decadal variations on a common trend. Additionally, a reconstruction from New Zealand suggests that changes in mid-latitude winds are
not purely zonal (Knudson et al., 2011). At higher latitudes, it has been proposed that the period from the 14th to the 19th centuries, corresponding to the LIA in the northern hemisphere, was characterized by a deepening of the Amundsen low (Kreutz et al., 2000), a weakening of the Antarctic High (Mayewski et al., 2004) and stronger meridional exchanges in the South Indian Ocean and southwest Pacific Ocean sectors of the Southern Ocean (Goodwin et al., 2004).

While a growing number of climate simulations spanning the past millennium have been achieved, most investigations of the model results have been focused on the best documented areas, and particularly the northern hemisphere continents (e.g. González-Rouco et al., 2003; Goosse et al., 2005; Jungclaus et al., 2010; Servonnat et al., 2010; Swingedouw et al., 2011). Only a few analyses were specifically devoted to the high latitudes of the southern hemisphere, a region where the majority of models displays large biases in simulations of the present-day mean state and variability (e.g. Arzel et al., 2006; Connolley and Bracegirdle, 2007; Lefebvre and Goosse, 2008; Monaghan et al., 2008b). For the last 500 years, Wilmes et al. (2012) show that the simulated temperatures in the southern hemisphere are strongly in instructive to also compare its results without data assimilation to the ones of the GCMs.

3. Methods

3.1. Simulations without data assimilation

We have first selected the simulations performed with two state of the art GCMs in the framework of PMIP3-CMIP5 (Paleoclimate Models Intercomparison Project phase 3 - Coupled Model Intercomparison Project phase 5): MPI-ESM-P and CCSM4 (Table 1). These simulations were chosen because they are the only ones that cover the full period 850–2000 and were available on the CMIP5 data portal at the time of the analysis. Two other experiments (performed with the GISS and MIROC models) were also available for the period 850–1850 but not for the period 1850–2000 for the same simulation. The MIROC run displays a surface temperature drift over the southern high latitudes of more than 3 °C in 1000 years, which is probably related to the experimental set up and not to a real feature of the system; an additional reason to discard it from our analysis. A preliminary analysis of the GISS outputs shows that the diagnostics discussed in Sections 4, 5 and 6 that we were able to perform using available data from this model are at least in qualitative agreement with the results obtained using the other models. Additionally, a larger ensemble of GCM simulations over the period 1850–2000 following the CMIP5 protocol is briefly used in Section 4 (Table S1).

We also analyze simulations performed with a previous version of the MPI model (MPI-ESM, Jungclaus et al., 2010, Table 1) because a larger ensemble is available (8 simulations compared to the single simulations with MPI-ESM-P and CCSM4) as well as experiments with the model of intermediate complexity LOVECLIM (Goosse et al., 2010). LOVECLIM has simpler dynamics and a coarser resolution than the GCMs. Its atmospheric component is ECBilt2 (Opsteegh et al., 1998), a quasi-geostrophic model with T21 horizontal resolution (corresponding to about 5.6° by 5.6°). ECBilt has no stratospheric dynamics. It is thus not possible to analyze tropospheric/stratospheric interactions, in particular the circulation changes in the troposphere resulting from modifications in stratospheric ozone concentration or in response to large volcanic eruptions (Goosse and Renssen, 2004). CLIO, the ocean component (Goosse and Fichefet, 1999), is a general circulation model with a horizontal resolution of 3° by 3°. A simple vegetation model (VECODE, Brovkin et al., 2002) is also activated in the configuration selected here, at the same resolution as in ECBilt. Because of those simplifications, LOVECLIM is much faster than the GCMs. Larger ensembles can thus be launched and simulations with data assimilation are possible (see Section 3.2). We consider it therefore instructive to also compare its results without data assimilation to the ones of the GCMs.

All the experiments are driven by natural (solar, volcanic and astronomical) and anthropogenic forcings (greenhouse gases, various aerosols, land use). MPI-ESM-P, CCSM4 and LOVECLIM follow the PMIP3-CMIP5 protocol (Schmidt et al., 2011). In particular they all use relatively modest changes in solar irradiance. The MPI-ESM forcing differs on some aspects because the simulations were launched before the CMIP5/PMIP3 protocol was formulated. The resulting changes for the first ensemble are small (E1, 5 simulations) compared with the other simulations and likely have a weak impact on the results. In the second ensemble (E2, 3 simulations), a larger scaling was applied to the solar forcing (an increase of 0.25% from the Maunder Minimum (1647–1715) to today in E2 compared to 0.10% in E1), leading to slightly larger temperature changes during the past millennium compared to the E1 ensemble (Jungclaus et al., 2010). In all those simulations, for simplicity, we will define the Antarctic surface air temperature as the temperature south of 70° S. We have checked in some models that computing the exact temperature over the continent does not make a significant difference for the type of analysis discussed here.

3.2. Data assimilation method

The data assimilation method is based on a particle filter (e.g., van Leeuwen, 2009) using the implementation described in Dubinkina et al. (2011). The experimental design is identical to the one applied in two recent studies (Goosse et al., 2012a, 2012b), albeit applied here to a new dataset of proxy records. Hence, only a brief description of the methodology is given here. Starting from an ensemble initialized for the year 851, 96 parallel simulations (called ‘particles’ or ensemble members) are propagated in time by

<table>
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<tr>
<td>CCSM4</td>
<td>National Center for Atmospheric Research, USA</td>
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<td>~0.6° × 0.9°</td>
<td><a href="http://www.cesm.ucar.edu/models/">http://www.cesm.ucar.edu/models/</a></td>
</tr>
<tr>
<td>MPI-ESM-P</td>
<td>Max Planck Institute for Meteorology, Germany</td>
<td>T63 (~1.875° × 1.875°)</td>
<td>~1° × 1°</td>
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</tr>
<tr>
<td>MPI-ESM</td>
<td>Max Planck Institute for Meteorology, Germany</td>
<td>T31 (~3.75° × 3.75°); Resolution ranging from 22 km to 350 km</td>
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<td>Jungclaus et al. (2010)</td>
</tr>
<tr>
<td>LOVECLIM</td>
<td>Université de Louvain, Belgium</td>
<td>T21 (~5.6° × 5.6°),</td>
<td>3° × 3°</td>
<td>Goosse et al. (2010)</td>
</tr>
</tbody>
</table>
the climate model. After one year, the likelihood of each particle is estimated as a function of the difference between the proxy-based reconstruction (see Section 3.3) and the simulated temperatures. The particles are then resampled according to their likelihood, i.e. to their ability to reproduce the signal derived from the proxy records. The particles with low likelihood are stopped, while the particles with a high likelihood are copied a number of times proportional to their likelihood in order to keep the total number of particles constant throughout the period covered by the simulations. A small noise is added to each copy to obtain different time developments for the following year. The entire procedure is repeated sequentially every year until the final year of calculation.

3.3. Data used in the data assimilation experiments

In the data assimilation experiments, LOVECLIM is constrained by a few summer and winter mean temperature reconstructions at mid latitudes (from tree rings and lake sediments) and by ice core records (Fig. 1, Table 2). As the high and mid latitude climates are coupled by large-scale circulation changes, we integrate the mid latitude proxy records to provide an additional constraint on the system. For all the regions, we selected among the available time series (see Neukom and Gergis, 2012) those that have at least a decadal resolution and which cover at least the period from the 14th century to the beginning of the 20th century. Additionally, we request for the mid-latitudes records a significant correlation over the period 1911–1990 with instrumental data within a search radius of 1000 km, at 5° × 5° spatial resolution, in order to keep only the series that are likely to reflect a large-scale temperature signal. The significance was tested on linearly detrended instrumental and proxy data, taking first order autocorrelations into account. We then scaled the proxy records to the mean and standard deviation of the instrumental data over the period 1911–1990.

For Antarctica, we have only selected isotope measurements in ice core records (Table 2). As the instrumental records are too short to provide a reliable local calibration, we assume that the isotope composition of the ice is related to the local temperature which is then derived here from δ18O measurements using a δT spatial slope of 0.8‰/°C (Masson-Delmotte et al., 2008), corresponding to the theoretical slope expected from the distillation of a given air mass. Temporal isotope-temperature gradients have been investigated at seasonal scale from observations (e.g. Law Dome, Dome F) and at decadal or longer time scale using isotopic atmospheric general circulation models (Schmidt et al., 2007; Sime et al., 2008). These studies generally point to a temporal gradient that appears systematically lower than the spatial slope, typically between 0.3 and 0.8‰/°C, due to covariance between precipitation and temperature, and possibly moisture origin; however, no study has been explicitly dedicated to the isotope-temperature slope at the multi-decadal time scale, relevant for the last millennium climate. In this framework, the slope of 0.8‰/°C appears as a conservative value which may lead to an underestimation (by a factor up to 2) of the temperature changes. Besides, the 0.8‰/°C slope is consistent with an independent estimate derived from bubble number-density measurements in West Antarctica (Fegyveresi et al., 2011). In order to reduce the non-climatic noise, a 10-year average has been applied to the isotope records before assimilation. Borehole temperature records and oceanic records, which have lower resolution, are not used in data assimilation.

For data assimilation, we need an estimate of uncertainty on temperature reconstructions. Here, our basic working hypothesis is that data error is assumed to be 0.5 °C and constant in time and space. This small uncertainty value is probably optimistic, especially when considering uncertainties in isotope-temperature relationships, but we selected it as a working hypothesis to provide a clear constraint on the model with data assimilation compared to the simulation without it and to check if the signal recorded in the proxies is compatible with model physics. As shown

![Fig. 1. Location of the proxy records used in the simulation with data assimilation on the atmospheric grid of LOVECLIM.](image-url)
in previous studies, our results are not extremely sensitive to this value and changing it by 50% does not have a major impact on the conclusions derived from our experiments with data assimilation (Goosse et al., 2012b). In addition, we did not take into account any error on the dating of the records as we will mainly focus our analyses on multi-decadal to multi-centennial time scales.

4. Model response to the forcing

As simulated for the northern hemisphere (e.g. González-Rouco et al., 2003; Goosse et al., 2005; Jungclaus et al., 2010; Servonnat et al., 2010; Swingedouw et al., 2011), the surface temperature over Antarctica displays a long term cooling trend between the beginning of the millennium and the mid-19th century in all the models, followed by a large warming until present-day (Fig. 2a). This recent warming is also obtained in the larger ensemble of ‘historical’ simulations covering the years 1850–2000 with a warming between 0.5 and 1.5 °C over that period in the different models (Fig. 3), confirming the results of Monaghan et al. (2008b) from a previous, smaller ensemble of simulations.

The delayed MCA described in Goosse et al. (2004), who were using an earlier version of LOVECLIM, is less clear in the new simulations performed with this model. The processes responsible for this different behavior have not been analyzed in detail but it is likely due to the smaller solar forcing applied in the current experiments, which leads to weaker temperature changes in the North Atlantic (Goosse et al., 2012b) and thus a smaller signal advected to the Southern Ocean compared to the earlier study. This feature also appears to be model dependent as it is not found in MPI-ESM-E2 despite the stronger solar forcing selected for this ensemble.

The magnitude of the pre-industrial decreasing temperature trend is small in LOVECLIM, in MPI-ESM-P and in MPI-ESM-E1. In these simulations, the difference between the warmest and coldest 50yr intervals reaches at most 0.2 °C–0.3 °C, with some periods of the 15th or 18th centuries being as warm as the early part of the simulation. On a multi-century timescale, the simulated difference in temperature between the warmest (950–1200) and the coldest intervals (1600–1850) is of 0.18 °C, 0.24 °C, 0.26 °C and 0.15 °C for LOVECLIM, MPI-ESM-E2, MPI-ESM-P and MPI-ESM-E1 respectively.

CCSM4 depicts the strongest pre-industrial trend, showing a difference of more than 0.7 °C between the coldest decades of the 14th, 18th and 19th centuries and the period 850–1200. The temperature difference between 950–1200 and 1600–1850 is 0.46 °C, i.e. around twice of that in the other models. This might be due to the climate sensitivity of CCSM4 (the climate sensitivity is

Fig. 2. a) Simulated annual mean near-surface air temperature anomaly southward of 70°S in LOVECLIM (red, mean of 10 simulations), MPI-ESM-P (orange), CCSM4 (dark blue), MPI-ESM-E1 (light blue, mean of 3 simulations) and MPI-ESM-E2 (magenta, mean of 5 simulations). b) Simulated annual mean temperature anomaly southward of 70°S in LOVECLIM driven by one forcing at a time: greenhouse gas (red), solar (green), land-use (blue), volcanic (light blue), orbital (magenta). c) Simulated seasonal mean temperature anomaly southward of 70°S in LOVECLIM (mean over 10 simulations) for December–January–February (red), March–April–May (green), June–July–August (Blue), September–October–November (light blue). d) The anomaly of ice area in the southern hemisphere (10^6 km²), in LOVECLIM (red, mean of 10 simulations), MPI-ESM-P (orange), CCSM4 (dark blue), MPI-ESM-E1 (light blue, mean of 3 simulations) and MPI-ESM-E2 (magenta, mean of 5 simulations). A 21-year running mean has been applied to the time series. Reference period is 1850–1980. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
a measure of the response of the model to a change in the external forcing and is defined as the equilibrium temperature change simulated following a doubling of the CO₂ concentration in the atmosphere; with a value of 3.2 °C, the climate sensitivity is higher in CCSM4 than in some of the other models used here (e.g. Goosse et al., 2010a, b), while being well in the range of the other GCMs and of the various estimates available (Bitz et al., 2012).

In order to illustrate the role of the various forcings, additional simulations have been performed with LOVECLIM driven by one forcing only at a time (as analyzed in Crespin et al., in press, for the Arctic). The difference in annual mean temperature between the 950–1200 and 1600–1850 intervals in response to volcanic and astronomical forcings reaches 0.07 °C, and 0.04 °C, respectively, compared to 0.18 °C in the simulation including all the forcings (Fig. 2b). The response to land use change also leads to a difference of 0.04 °C between those two periods, underlining the non-negligible contribution of this mid and low latitudes forcing in polar areas, as also noticed for the Arctic (Crespin et al., in press). Those estimates should be considered as an order of magnitude only as they depend on the model and forcing time series selected but they show that volcanic, land-use and astronomical forcings can largely explain the pre-industrial cooling trend in the experiments performed, the solar forcing playing a small role. By contrast, the 20th century warming is mainly attributed to the greenhouse gas forcing. The role of ozone changes has not been investigated in LOVECLIM because of the model limitations mentioned in Section 3.1. Note that at seasonal scale, the contribution of the astronomical forcing is larger as discussed below.

In the simulations performed with all the models, the pre-industrial cooling trend is larger in Antarctica than at global scale. The ratio of the cooling between the periods 850–1200 and 1600–1850 in Antarctica and of the one at global average, which is a measure of the polar amplification of the temperature changes, is between 1.2 and 1.9. This is, however, much smaller than for the Arctic where the corresponding ratio equals 4.6 for LOVECLIM (which is known to have a strong polar amplification there, Goosse et al., 2010a, b) and between 2.42 and 2.85 for the other models.

Within its pronounced pre-industrial trend, CCSM4 displays a step-like cooling of about 0.3 °C between the first and the second half of the 13th century. In this case, only the 20th century warming allows Antarctic temperatures to reach a level encountered prior to the 13th century. In simulations performed with an earlier version of the model (CCSM3), Miller et al. (2012) attributed a similar abrupt cooling in the Arctic to the response of the model to volcanic forcing, amplified by sea-ice/ocean feedbacks (Zhang et al., 2011). The results of CCSM4 are coherent with a similar mechanism also operating in the Southern Ocean in this model version (Fig. 2d). In CCSM4, the difference in annual mean sea ice area between its maximum and minimum of the pre-industrial period is larger than 1.5 × 10⁶ km², which is much higher than in other models (corresponding values between 0.4 × 10⁶ km² and 0.9 × 10⁶ km²). The most pronounced decrease in sea ice concentration occurs in the Bellingshausen Sea (not shown). This is consistent with the larger temperature changes simulated by CCSM4 there as well as in West Antarctica (see Fig. 4).

Simulations driven by similar forcings do not display changes that are coherent in time, because of the strong magnitude of the internal variability in the high latitudes of the southern hemisphere (Fig. 2a). This internal variability can be dominant for specific seasons. As a consequence, the robust long term trend is difficult to identify for a given season in the individual simulations (not shown), except in CCSM4 which displays similar changes for each season and for the annual mean because of the larger signal. Performing the average over an ensemble, as in LOVECLIM, tends to smooth out the internal variability (Fig. 2c) and the resulting ensemble mean depicts a response to the forcing characterized by larger pre-industrial cooling trends in spring (SON) and summer (DJF). The solar energy received at the top of the atmosphere between October and mid January at high southern latitudes decreased by up to 9 Wm⁻² during the past millennium because of the influence of precession on seasonal insolation (see Fig. 1 of Schmidt et al., 2011). The cooling simulated in spring and summer is thus consistent with a delayed response to those local insolation changes, with temperatures lagging insolation by 1–2 months because of the thermal inertia of the system, as discussed in more details in Renssen et al. (2005) for the whole Holocene.

At low and mid-latitudes, the response to a forcing is generally larger over land than over the ocean. This is due to the thermal inertia of the ocean, which damps the transient changes as well as

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Fig. 3. Simulated annual mean near-surface air-temperature anomaly southward of 70° S in historical runs performed by 27 GCMs following CMIP5 protocol (for a definition of the model names see Massonnet et al., 2012 or Table S1) and in LOVECLIM without data assimilation. An 11-year running mean has been applied to the time series. The reference period is 1850–1880.
to some negative feedbacks, in particular those involving modifications in latent heat fluxes and in ocean-land exchanges (e.g. Boer, 2011). In some models, such as MPI-ESM (E1 and E2), larger changes are simulated over Antarctica compared to the Southern Ocean (Fig. 4) between the warmer (950–1200) and colder (1600–1850) periods analyzed above. In LOVECLIM, and MPI-ESM-P, the temperature difference is generally larger over land than ocean, but the absolute maxima are found over coastal seas, advocating for local amplification by simulated sea-ice ocean interactions. Consistent with the hypothesis proposing that the feedbacks involving sea-ice play a larger role in CCSM4 than in the other ones, the response in CCSM4 is also clearly larger over the ocean than over land.

In addition to the temperature changes, we have also analyzed the atmospheric circulation changes in the different models. Changes are weak and the inter-model spread particularly large. The majority of models display slightly higher pressure over Antarctica during the years 950–1200 than during the years 1600–1850 but these changes are not robust between the simulations (not shown).

Fig. 4. Difference in annual mean near-surface air temperature (°C) between the periods 950–1200 and 1600–1850 in all the models.
5. Analysis of the simulations with data assimilation and model data comparison

At the grid-scale, as expected, the simulation with data assimilation in LOVECLIM generally follows relatively well the proxy data used to constrain the model evolution (Fig. 5). The variability is lower in the simulations, but this is a classical bias in order to reduce the model error compared to the proxy-based reconstructions (Goose et al., 2010a, b; Annan and Hargreaves, 2012; Goose et al., 2012a, b). The difference between model and reconstructions is generally of the order of a few tenths of a degree, which is perfectly acceptable knowing the uncertainty of proxies.

There are, however, exceptions. The model is not at all able to follow the long term temperature decrease deduced from the Law Dome ice core isotope data. The agreement at multi-decadal scale is also weak for EDC and in Tasmania (and to a smaller extent for New Zealand). Because of its coarse resolution, LOVECLIM does not include land in the grid point corresponding to Tasmania. We are thus considering a reconstruction over land with model results over the ocean which displays more muted variations. New Zealand is represented by two grid-points in the model and is thus also strongly influenced by surrounding oceanic conditions.

Multiple processes can be responsible for the model-data mismatch at Law Dome and EDC. Temperature reconstructions based on isotope measurements suffer from uncertainties related to precipitation intermittency, changes in the moisture origin and trajectory, changes in the boundary layer structure (condensation versus surface air temperature), and changes in local elevation, although the latter is likely not dominant at the millennial timescale (e.g., Sime et al., 2009; Stenni et al., 2010; Sidall et al., 2012). Processes potentially responsible for a long term bias may have only a weak impact on the short term, making any validation using recent thermometer data only partial.

On the other hand, adequately representing high latitude climate dynamics remains a challenge for climate models, as illustrated by the large biases in the simulations of the present state of the system (e.g., Arzel et al., 2006; Connolley and Bracegirdle, 2007; Lefebvre and Goose, 2008; Monaghan et al., 2008b). Global climate models do not have the adequate resolution to reproduce well the local topography and thus the atmospheric conditions leading to snow deposition in coastal sites such as Law Dome. This problem seems particularly important for coarse resolution models as LOVECLIM whose grid cell corresponding to Law Dome is going up to 500 km inland and is thus certainly not representative of coastal conditions.

Data assimilation looks for the best compromise between model physics and observations in order to maximize the likelihood of the model state. LOVECLIM is not able to reproduce the temperature trend deduced from Law Dome ice core isotope measurements, while it can be brought in agreement with the majority of records. Note that the model-data mismatch for Law Dome is not restricted to LOVECLIM, as none of the models used here simulates a long term cooling over the period 850–2000 at the location corresponding to Law Dome (see Section 5).

Averaged over Antarctica, the simulation with data assimilation displays a similar multi-centennial development as the simulation with LOVECLIM without data assimilation (Fig. 6a, b). Differences are only seen during specific, relatively short periods. For instance, the early 15th century and the 1940s are warmer by 0.1–0.2 °C in the simulations with data assimilation while the late 10th century and the late 20th century are colder than without data assimilation by about the same amount. As a consequence, the data assimilation slightly reduces the recent warming trend, which, however, remains larger in the model than in the instrumental temperatures, in particular at the end of the record. Previous simulations have shown that the recent trend is well reproduced if the model is constrained by instrumental data using data assimilation (Goose et al., 2009). We therefore suspect that the model-data mismatch for the recent trend could be caused by the scarcity of recent proxy records, as many ice cores were drilled before the 1990s and their records therefore do not cover the last decades. We will thus not discuss the most recent changes here (since 1990). The simulations capture correctly the warming between the 1960s and 1980s present in direct measurements and in the reconstructions of Schneider et al. (2006), which is based on data from Law Dome, DML (for those two records, an earlier version was used compared to the one included here), Siple Station (75° 92 S 84° 10 W), and two shallow west Antarctic cores (79° 38 S 111° 23 W and 77° 68 S 123° 99 W). Before 1950, the simulations with data assimilation display similarities with the Schneider et al. (2006) reconstruction on decadal scale, such as the warming around 1900 and 1940. Altogether, the simulated temperature trend since the 19th century is, however, much larger in the simulations with LOVECLIM, and in the simulation performed by the GCMs (Fig. 3), than in the reconstruction of Schneider et al. (2006). For instance, the difference between the years 1850–1880 and 1960–1990 reaches 0.18 °C in Schneider et al. (2006) and 0.51 in our simulations with data assimilation.

Compared to the simulations without data assimilation in LOVECLIM, the data assimilation reduces sea ice retreat over the last decades, which is consistent with the smaller warming over that period (Fig. 6c). Besides, the sea ice area is smaller in the simulation with data assimilation during nearly the whole period 850–1500, with differences reaching 0.5·10^6 km^2. As a consequence, the simulated sea ice area is similar at the end of the 20th century and around 900, 1220, 1350 in the simulations with data assimilation while the recent decades are characterized by a clear minimum in LOVECLIM without data assimilation as well as in the majority of the GCMs simulations (Fig. 2d). Unfortunately, those simulated long term changes in sea ice area could not be assessed from observations because of the lack of high resolution sea ice data for the past millennium.

For the pre-industrial period, as no strong large-scale signal comes out at the scale of the continent, it is instructive to analyze the spatial distribution of the changes. Because of the data constraints, the structure of the changes between 950–1200 and 1600–1850 is different in LOVECLIM with assimilation (Fig. 7a) compared to the simulations without data assimilation (see Fig. 4a). Consistently with the ice core data, the period 950–1200 is much warmer close to DML. The area close to Siple and Taylor domes is colder, while slightly warmer conditions are obtained around EDC and Law Dome. In the model, this is associated with a higher geopotential height (and equivalently higher surface pressure) over the Bellingshausen and Amundsen Seas and a lower one over the continent during the period 950–1200 than during the period 1600–1850 (Fig. 7c). Those circulation changes are consistent with the interpretation of the circulation changes of Kreutz et al. (2000) based on the aerosol transport to Siple Dome. This circulation pattern induces northerly winds in the Ross Sea leading to a large local warming, in agreement with the reconstructed warming between 1140 and 1280 (Bertler et al., 2011). Slightly warmer conditions are also simulated on the western side of the Antarctic Peninsula during the period 950–1200, consistent with local reconstructions (Shevenell et al., 2011). These two data sets, which were not used in our data assimilation experiment, therefore provide an independent support for our results. Temperature changes simulated by our coarse resolution model appear, however, much weaker than the reconstructed ones. We finally investigated simulated changes in westerlies, showing the lack of any significant variation in strength or position, in contrast with...
Fig. 5. Comparison of the proxy-based reconstructions (black) used to constrain model evolution with the annual mean temperature anomalies in the simulation with data assimilation at the corresponding location (red). A 21-year running mean has been applied to the time-series. The reference period is 1850–1980. All the proxies available for New Zealand and South America have been averaged in the bottom two panels. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
interpretation of mid latitude observations (see Section 2). This results should, however, be taken with caution as it might be related to the low number of proxies used to constrain model results at mid-latitudes or to the weak tropical extra-tropical connection in LOVECLIM.

It is out of the scope of our study to analyze in more details the spatial structure of the changes. The large uncertainty in model results over Antarctica and the limited amount of data covering the past millennium preclude any definitive conclusion on the dominant characteristics of the variability at the regional scale and on the mechanisms ruling them. The spatial structure in the model with data assimilation crucially depends on the choice of the proxies to constrain the model results. Nevertheless, we note that, at multidecadal time scale, some records (e.g. EDC versus Siple and Taylor Domes) are clearly anti-correlated. Regional anti-correlation is also depicted in MPI-ESM-P between the locations of Siple Dome and B25 (see also Fig. 10c below). For instance, when comparing a 50-year period of the first half of the millennium with the reference period 1850–1980, warm anomalies are simulated in some coastal regions of both East and West Antarctica while a band of cold anomalies is depicted from the Weddell Sea to Wilkes Land (Fig. 8).

6. A composite of the ice core records

Fig. 5 shows the spread between individual ice core records. When plotting all the time series on the same graph and applying a stronger temporal filter, the diversity of the temperature changes appears even more clearly (Fig. 9a). For instance, EDC, Taylor, Siple and Tallice reconstructions all display a clear warming after 1800. DML, B25 and Law Dome reconstructions show maximum temperatures before the 20th century; around the beginning of the period of interest for DML and Law Dome, in the 14th century for B25. In contrast to the other 2 records, DML reconstruction also shows a warming during the 20th century, which is, however, not exceptional in the framework of the last 1150 years. Ice cores located in East Antarctic regions relatively close to the coast, in the Atlantic and Indian sectors, do not depict any remarkable warming during the 20th century. The density of the cores is, however, too low to draw a boundary between the behavior of the different regions.

The coherency between the different regions is larger in models than in the reconstructions based on ice core records, even in the two models that have the highest resolution among the ones we analyzed here (Fig. 10a, c). At the multi-decadal scale, models do simulate contrasted variability in some regions as discussed in Section 5. However, multi-centennial trends appear consistent over the whole continent. As a consequence, the average of model results at those seven ice core sites is very close to the simulated Antarctic mean temperature (Fig. 10b, d).

This indicates that, in the models, a sample of 7 cores is enough to have a very good measure of the mean over the continent. We therefore produce an estimate of the temperature change in Antarctica, based on the average of the 7 ice core records (Fig. 9b). This rough first-order reconstruction, referred to below as the Antarctic Isotopic Temperature composite (AIT), is based on the simplest possible technique. We consider nevertheless this exercise to be useful since more complex methods may suffer from the very small overlap between thermometer measurements and the majority of the long, high-resolution records selected here. This new reconstruction of Antarctic temperatures shows persistent, large multi-decadal variability during the whole record, a small cooling trend during the preindustrial period and a modest warming mainly between 1800 and 1970. Within uncertainties, the recent decades are not the warmest decades of the millennium, the absolute maximum of the AIT composite being found in the 15th century. Over the last 70 years, our AIT composite is in very good agreement with the reconstruction of Schneider et al. (2006) and
instrumental observations over the last decades (shown in Fig. 6a). This supports our isotope-temperature calibration using a $\delta^4/\Delta T$ spatial slope of $0.8 \pm 0.1 \degree C$. Our reconstruction, however, diverges from the one of Schneider et al. (2006) prior to 1930, a period for which not enough direct observations are available in Antarctica to estimate their respective skill.

As expected, the reconstruction based on the 7 ice core measurements is in good agreement with the simulations with data assimilation. This indicates that the AIT composite is compatible with the model physics and the forcing applied in LOVECLIM. In particular, the difference between the years 1850–1880 and 1960–1990 reaches $0.45 \degree C$ in the AIT composite (compared to $0.51 \degree C$ in LOVECLIM with data assimilation, see Section 5). The AIT composite is also in reasonable agreement with the results of the other models (see Fig. 2a). The long term trends are, however, weaker in the AIT composite. The difference between the intervals 950–1200 and 1600–1850 which were the warmest and coldest in the models is only of $0.03 \degree C$ in the AIT composite, compared to $0.1 \degree C$ in LOVECLIM with data assimilation and more than $0.15 \degree C$ in the different simulations without data assimilation (see Section 4).

Besides, the multi-decadal variability of the AIT composite is higher than the one of the models (except CCSM4), as already noticed in Section 5 for LOVECLIM. The standard deviation of the AIT composite has a value of $0.22 \degree C$ after a 21-year smoothing over the period 850–1850, compared to $0.17 \degree C$, $0.14 \degree C$, $0.17 \degree C$, $0.11 \degree C$ and $0.23 \degree C$ for MPI-ESM-P, MPI-ESM-E1, MPI-ESM-E2, LOVECLIM with data assimilation and CCSM4, respectively. The agreement between the model and reconstructions on the magnitude of the interdecadal to multi-centennial variability is, however, much better at the continental scale than for the individual records (compare Figs. 3, 8 and 9).

The extrapolation of the conclusion derived from model results to justify our average of the proxy-based reconstruction should, however, be taken with caution. Indeed, as mentioned above, the spatial coherency is larger in models than within the ice core records. Furthermore, we have not taken into account that ice core records are composed of a local temperature signal, but also non-temperature noise caused by deposition or post-deposition
processes. Water stable isotope records from ice cores archive climatic information at times when precipitation occurs. Precipitation intermittency and moisture source changes may thus affect temperature reconstructions based on ice core records of precipitation isotopic composition (Masson-Delmotte et al., 2011). Biasing effects (seasonal and synoptic), reflecting warmer temperatures during snowfall, may also differently affect inland and coastal Antarctic sites and could explain the large spread between

Fig. 9. a) Temperature anomaly (°C) reconstructed from different ice core records: DML (light green), Taldice (yellow), Siple (red), Law Dome (magenta), B25 (violet), EDC (black), Taylor (dark green). b) Mean of the temperatures reconstructed from all the ice core records (AFT composite, light blue), a previous reconstruction based on ice core data (dark blue, Schneider et al., 2006) and the mean temperature anomaly southward of 70° S in the simulation with data assimilation (green). Additionally, the 7 estimates obtained by performing the mean over only 6 of the 7 records are also displayed in grey to illustrate the uncertainty of the reconstruction. A 101 year running mean and a 21-year running mean have been applied for panel a and b, respectively. Reference period is 1850–1980. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Fig. 10. a) Temperature anomaly (°C) simulated in CCSM4 at the location of the different ice core records: DML (light green), Taldice (yellow), Siple (red), Law Dome (magenta), B25 (violet), EDC (black), Taylor (dark green). b) Mean of the temperature anomalies simulated at the location of all the ice core records (red) and the mean temperature southward of 70° S in CCSM4 (blue). A 51 year running mean and a 21-year running mean have been applied for panel a and b, respectively. c and d, same as a and b for the MPI-ESM-P model. Reference period is 1850–1980. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
ice core records. A single model grid–point temperature is representative of a larger area, because of the relatively coarse model resolution, than an individual proxy record. Consequently, the mean over seven model grid points is certainly a much better estimator of the average temperature simulated over Antarctica, compared to the mean of seven temperature estimates derived from ice core stable isotope records. Nevertheless, the results remain robust if we choose any subset of 6 of the 7 cores, giving some confidence that the number of records is sufficient to resolve a common climate signal (Fig. 9b).

7. Conclusions

Despite the uncertainties on proxy reconstructions and models, they consistently rule out large (>0.5 °C), homogeneous pre-industrial Antarctic temperature variations during the last millennium, in contrast to the coherent variability displayed on longer timescales (e.g., Masson et al., 2000; Masson-Delmotte et al., 2011). The spatial structure of the changes appears to be complex, with nearby regions such as the Ross Sea and the nearby Siple and Taylor Domes experiencing opposite variations even at the multi-century time-scale. Consequently, we could not detect using the available information a transition at the scale of Antarctica equivalent to the one proposed in the northern hemisphere between the MCA and LIA. Large regional differences also appear on shorter time scales (e.g. decadal). During some periods, contrasted behaviors emerge between East Antarctica and West Antarctica, as noticed for the recent past but such distinction does not necessarily hold at centennial time-scale during the past millennium.

The signal simulated by the model using data assimilation is broadly consistent with proxy-based reconstructions, with local exceptions. The data constraint does not induce significant shifts in model state during centuries in the regions where the proxy data are assimilated. This contrasts with the earlier studies devoted to the northern hemisphere (Goosse et al., 2012a,b). In those previous studies, the data were showing a robust large scale structure, related to the LIA-MCA transition, that has a too weak amplitude in LOVECLIM without data assimilation. Model results were thus adjusted to better fit the data thanks to the assimilation technique. As such robust large-scale pattern is not present in the data selected here, no corresponding shift is associated with the data assimilation.

As a consequence of the absence of a large common signal between the different regions, the mean temperature over the continent only explains a small part of the variance of the system. In models, it mainly displays the response to the natural and anthropogenic forcing applied, characterized by a cooling trend during the preindustrial period followed by a warming over the last 150 years. This picture is qualitatively similar to our Antarctic isotopic temperature composite obtained by simply averaging the reconstructions based on the 7 ice core records. Nevertheless, this reconstruction of the mean Antarctic temperature from ice core data can be sensitive to the choice and the representativeness of the records and should thus be considered as a rough first-order estimate only that should be refined in the future. Model-data comparisons should be further investigated, for instance using precipitation weighted temperature or isotopically enabled models. The AIT composite is not strongly affected if we do not include one of the seven records but selecting only a few time series in the average could lead to a clearly different reconstruction. The agreement with model results at the centennial scale, however, provides an encouraging cross validation of two uncertain sources of information.

The AIT composite is in good agreement with another reconstruction (Schneider et al., 2006) and with estimates based on direct observations over the last decades. The reconstructed warming of about 0.5 °C over the last 150 years is, however, about two times larger than the one obtained by Schneider et al. (2006). This 0.5 °C warming obtained in the AIT composite is very close to the value obtained in our simulations with data assimilation and is well in the range simulated by current general circulation models driven by realistic forcings. It is also consistent with some additional reconstructions using ocean and ice core data and with several borehole temperature measurements. It is, however, not clearly detected in many isotope records from ice cores, including some of our data and others used in Schneider et al. (2006) as well as more recent ones (e.g. Sinclair et al., 2012). Again, isotopic enabled models may allow to understand the processes at play.

Estimating precisely the magnitude of the warming over the last 150 years compared to the pre-industrial climate is of paramount importance for our understanding of the response of Antarctica to external forcings. This is also essential for the models validation as reproducing the temperature changes over that period is probably the most crucial test of models before using them to study future changes over Antarctica. Our approach focusing on ice cores covering at least several centuries is different from the one of Schneider et al. (2006) that is devoted to the last 200 years only. We thus have not taken into account all available information spanning the recent decades. Obtaining additional reliable series of Antarctic temperatures for the whole millennium, including the recent decades, should then be considered as a central objective, requiring additional records to have a reasonable sampling of the spatial variability of the system. Such a synthesis would also provide essential information on the spatial structure of the changes over this period. This is challenging but also the only way to understand the climate dynamics at decadal to centennial timescales and to test the ability of models to adequately represent it. The dating, aiming to annual resolution, and the interpretation of the proxies should also be improved to have more reliable reconstructions and more sound model-data comparisons. The latter should include the comparison of different types of proxies in order to have a precise estimate of the uncertainties as well as a better documentation of the isotope temperature relationship on a variety of time scales, for instance using borehole data when possible and direct proxy modeling.

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