Global and Planetary Change 84-85 (2012) 35–47

Contents lists available at ScienceDirect

Global and Planetary Change

journal homepage: www.elsevier.com/locate/gloplacha

The medieval climate anomaly in Europe: Comparison of the summer and annual mean signals in two reconstructions and in simulations with data assimilation

Hugues Goosse [a,*], Joel Guiot [b], Michael E. Mann [c], Svetlana Dubinkina [a], Yoann Sallaz-Damaz [a]

[a] Université Catholique de Louvain, Earth and Life Institute, Georges Lemaître Centre for Earth and Climate Research, Chemin du Cyclotron, 2, B-1348 Louvain-la-Neuve, Belgium
[b] CEREGE, Aix-Marseille Université/CNRS, Europole Méditerranéen de l'Arbois, Aix-en-Provence, France
[c] Department of Meteorology and Earth and Environmental Systems Institute, Pennsylvania State University, University Park, USA

ARTICLE INFO

Article history:
Received 7 January 2011
Accepted 6 July 2011
Available online 20 July 2011

Keywords:
Europe
Last Millennium
seasonal climate signal
atmospheric circulation
data assimilation

ABSTRACT

The spatial pattern and potential dynamical origin of the Medieval Climate Anomaly (MCA, around 1000 AD) in Europe are assessed with two recent reconstructions and simulations constrained to follow those reconstructions by means of paleoclimate data assimilation. The simulations employ a climate model of intermediate complexity (LOVECLIM). The data assimilation technique is based on a particle filter using an ensemble of 96 simulations. The peak winter (and annual mean) warming during the MCA, in our analyses, is found to be strongest at high latitudes, associated with strengthened mid-latitude westerlies. Summer warmth, by contrast, is found to be greatest in southern Europe and the Mediterranean Sea, associated with reduced westerlies and strengthened southerly winds off North Africa. The results of our analysis thus underscore the complexity of the spatial and seasonal structure of the MCA in Europe.
© 2011 Elsevier B.V. All rights reserved.

1. Introduction

Europe is one region of the world for which a large number of proxy records are available spanning the past millennium (e.g. Lamb, 1965; Pfister et al., 1998; Klimenko et al., 2001; van Engelen et al., 2001; Shabalova and van Engelen, 2003; Charman, 2010; Guiot et al., 2010). The density of proxy records is particularly great in the Alps (e.g., Büntgen et al., 2005; Mangini et al., 2005; Büntgen et al., 2006; Corona et al., 2010) and in Fennoscandia (e.g. Briffa et al., 1992; Hiller et al., 2001; Tiljander et al., 2003; Kremenetski et al., 2004; Grudd, 2008; Seppa et al., 2009; Gunnarson et al., 2010). These proxy records generally indicate relatively warm conditions during Medieval Time, i.e. between roughly 850 and 1250 AD. This is particularly clear for summer as the majority of the proxies represent summer or growing season (April to September) temperature. However, the few direct estimates of winter and annual mean changes also indicate relatively mild conditions (e.g., Pfister et al., 1998; Van Engelen et al., 2001; Shabalova and van Engelen, 2003; Tiljander et al., 2003).

This has been one of the bases for the quite loose definition of the so-called Medieval Warm Period or Medieval Climate Anomaly (MCA, e.g., Lamb, 1965; Hughes and Diaz, 1994; Bradley et al., 2003). Nevertheless, Medieval warmth was certainly not continuous, nor was it spatially synchronous (e.g. Hughes and Diaz, 1994; Bradley et al., 2003; Goosse et al., 2005; Guiot et al., 2010). In particular, several records indicate cold conditions between 1050 and 1150 AD (e.g. Büntgen et al., 2006; Grudd, 2008; Corona et al., 2010) with temperatures similar to those observed later, during what is often referred to as the Little Ice Age (LIA, roughly between 1400 and 1850).

The generally mild European conditions during the MCA could potentially be the consequence of a prolonged positive anomaly in external radiative forcing. Medieval Times are characterised by high Total Solar Irradiances (TSI) around 900–1000 and 1050–1250 (Bard et al., 2000; Muscheler et al., 2007; Delaygue and Bard, 2010) which should have contributed to observed surface warming. Major explosive volcanic eruptions cool the surface, in particular in summer, by reducing the incoming shortwave radiative heating (Robock, 2000; Fischer et al., 2007). The relative absence of major eruptions during the MCA relative to the later LIA (Crowley et al., 2003; Gao et al., 2008) thus also likely contributes to the level of warmth at that time. Additionally, Europe has been subject to major land use changes during the past millennium (Goudie, 1993; Pongratz et al., 2008; Kaplan et al., 2009). As deforestation increases the surface albedo (biogeophysical effect), this induces a general cooling tendency in Europe since Medieval Times (Brovkin et al., 2006; Goosse et al., 2006a; Pongratz et al., 2009). Deforestation also affects the surface characteristics governing the evapotranspiration and the heat and momentum exchanges between the atmosphere and the land surface. However, the net effect of those changes is usually harder to evaluate than the one of the albedo (Matthews et al., 2004; Feddema et al., 2005; Pitman et al., 2009). Finally, deforestation also has an impact on the global atmospheric CO_2 concentration (biogeochemical effect) inducing a large-scale warming that compensates in some regions for the cooling caused by the biogeophyscial effect (Pongratz et al., 2010).

* Corresponding author. Tel.: +32 10 47 32 98; fax: +32 10 47 47 22.
E-mail address: hugues.goosse@uclouvain.be (H. Goosse).

0921-8181/$ – see front matter © 2011 Elsevier B.V. All rights reserved.
doi:10.1016/j.gloplacha.2011.07.002

In addition to these predominantly thermodynamic effects, volcanic and solar forcing can profoundly influence the large-scale atmospheric circulation. For instance, there is a tendency for the positive phase of the North Atlantic Oscillation (NAO) in the winter following a major volcanic eruption. Positive solar radiative forcing, similarly, is associated with the positive phase of the NAO. These effects imply a relatively complex regional spatial response, with either an amplification or damping of the direct radiatively-forced temperature responses depending on the region (Robock, 2000; Shindell et al., 2001, 2003; Fischer et al., 2007).

Finally, purely internal variability of the climate system can potentially play a dominant role in the climate changes observed at regional and continental scales during the past millennium (Goosse et al., 2005; Servonnat et al., 2010). The positive phase of the NAO leads to relatively warm winters over much of Europe due to the associated stronger onshore flow of warm maritime air off the Atlantic Ocean (Hurrell, 1995; Wanner et al., 2001; Trouet et al., 2009). Since much of Europe is under the influence of westerly winds, any change in sea surface temperature over the Atlantic, ocean (e.g. due to changes in oceanic circulation) can impart a strong imprint on temperatures in Europe through downstream advection (e.g. Sutton and Hodson, 2005; Pohlmann et al., 2006; Semenov et al., 2010). It has even been argued that internal variability alone might explain the observed temperature variations in Europe during pre-industrial times (Bengtsson et al., 2006). External forcing is however – at least for certain periods of the past millennium – a very likely contributor to the observed climate changes (e.g., Fischer et al., 2007; Hegerl et al., 2007, 2011).

As briefly summarised above, the mechanisms that might potentially explain the observed temperature changes in Europe over past centuries have been investigated in several studies. Those studies included analysis of paleoclimate reconstructions and the use of various types of climate models. Comparing model results with reconstructions to infer the dominant processes responsible for any particular observed change in climate is not straightforward however. Due to the chaotic nature of internal climate variability, it is exceedingly unlikely that any coupled climate model will be able to reproduce the precise realisation of past climate, i.e. will be able to fortuitously capture the actual phases of the various internal modes of climate variability, influencing for instance temperature patterns, without additional real-world constraints. Such constraints can be provided by the climate observations themselves. With the possible exception of a specific very strong external forcing scenario (e.g. Timmreck et al., 2009), models and data can compare only in a statistical sense (e.g.; Brewer et al., 2006). This is a quite strong limitation if one is interested in understanding the observed changes associated with a particular time period such as the MCA.

Complementary information can be obtained using paleoclimate data assimilation, combining model results and proxy climate observations to obtain an optimal estimate of the past states of the system (e.g., Widmann et al., 2010). Since data assimilation constrains the model to follow observations within their uncertainties, it provides an estimate of the full climate state, i.e. a suite of simulated atmospheric, oceanic and land variables, that is most compatible with the information in the proxy data. The method therefore allows for the investigation of specific hypotheses regarding the role of large-scale climate dynamics in explaining the paleo-observations (e.g.; van der Schrier and Barkmeijer, 2005; Crespin et al. 2009; Widmann et al., 2010).

In the present study, we use a particle filter data assimilation method (van Leeuwen, 2009). Two different recent surface temperature reconstructions (Mann et al., 2008, 2009; Guiot et al., 2010) are assimilated into the climate model LOVECLIM (Goosse et al., 2010a). One of the two reconstructions (Mann et al., 2009) estimates annual mean temperatures patterns across the globe for the last 1500 years, while the second (Guiot et al., 2010) estimates warm season (i.e. growing season/summer) temperatures in Europe over the past 1400 years. Use of these dual reconstructions thus allows for an analysis of the seasonal characteristics of European temperature changes and their relationship with larger-scale climate changes over the past millennium.

The model, the proxy-based reconstructions, and the data assimilation method are briefly presented in Section 2. Section 3 describes and analyses the simulations of European climate during the MCA interval. We discuss our results in Section 4 before presenting our concluding remarks.

2. Experimental setup

In order to perform the simulations with data assimilation that will be used to study European climate, we have to select a model (Section 2.1), a forcing scenario (Section 2.2), initial conditions (Section 2.3), a data assimilation method (Section 2.4), proxy-based reconstructions (Section 2.5), and a way to measure the agreement between model results and the reconstruction (Section 2.6). Simplifications and assumptions are required on each of those steps. Many of them are imposed by technical constraints: the model must represent sufficiently well the physics of the system but must also be fast enough to apply data assimilation techniques; the number of simulation in an ensemble should be of the order of 100 at most because of computer-time limitation although a larger number might be suitable to precisely characterise variability at all spatial scales; the two spatial reconstructions, for two different seasonal ranges, that are available for our period of interest are based on a limited number of proxy records and on methods that may have biases. In order to formulate precisely the problem, an uncertainty must thus be a priori associated to all the important assumptions, as described in each of the sub-sections below. Despite those assumptions and uncertainties, we will show that our simulations with data assimilation bring new results compared to simulations without data assimilation and contribute to improve our knowledge of European climate during the MCA. This will be evaluated a posteriori by ensuring the compatibility of the results with our working hypotheses and by analysing the useful information brought in Sections 3 and 4.

2.1. The climate model

LOVECLIM 1.2 is a three-dimensional Earth system model of intermediate complexity that includes representations of the atmosphere, the ocean and sea ice, the land surface (including vegetation), the ice sheets, the icebergs and the carbon cycle. Goosse et al. (2010a) provide a comprehensive description of the current model version as well as a brief comparison of the simulation results with the observed climate both for present-day conditions and key past periods.

The atmospheric component is ECBilt2, a quasi-geostrophic model with T21 horizontal resolution (corresponding to about 5.6° by 5.6°) and 3-level on the vertical (Opsteegh et al., 1998). The ocean component is CLIO3 (Goosse and Fichefet, 1999), which consists of an ocean general circulation model coupled to a thermodynamic-dynamic sea-ice model. Its horizontal resolution is of 3° by 3°, and there are 20 levels in the ocean. ECBilt–CLIO is coupled to VECODE, a vegetation model that simulates the dynamics of two main terrestrial plant functional types, trees and grasses, as well as desert (Brovkin et al., 2002). Its resolution is the same as of ECBilt. In order to reduce the computational time requirements of the model and because of the focus of our study, the ice sheets, iceberg and carbon cycle components are not activated in our simulations. As a consequence, we cannot compute the time development of the atmospheric CO_2 concentration interactively or take into account explicitly the biogeochemical impacts of land use changes as, for instance, done in Pongratz et al. (2010) and Jungclaus et al. (2010).

Download English Version:

https://daneshyari.com/en/article/6348413

Download Persian Version:

https://daneshyari.com/article/6348413

Daneshyari.com