Quantifying climate forcings and feedbacks over the last millennium in CMIP5

A.R. Atwood\textsuperscript{a}, E. Wu\textsuperscript{b}, D.M.W. Frierson\textsuperscript{b}, D.S. Battisti\textsuperscript{b}, J.P. Sachs\textsuperscript{a}

\textsuperscript{a} University of Washington, School of Oceanography, Seattle, WA 98195, USA

\textsuperscript{b} University of Washington, Dept. of Atmospheric Sciences, Seattle, WA 98195, USA

* Corresponding author. Tel: +1 206-694-3143; fax: +1 206-685-3351

E-mail address: aatwood@uw.edu
Abstract

The role of radiative forcings and climate feedbacks on last millennium climate change is quantified in the transient climate model simulations from the CMIP5/PMIP3 archive. Changes in the global energy budget over the last millennium are decomposed into contributions from radiative forcings and climate feedbacks through the use of the Approximate Partial Radiative Perturbation (APRP) method and radiative kernels. Global cooling occurs ca. 1200-1850 CE in the multi-model ensemble mean with pronounced minima ca. 1230-1290, 1450, 1600-1700, and 1820-1840 CE (corresponding with volcanically active periods) that are outside the range of natural variability. Average Northern Hemisphere temperature anomalies agree well with proxy-derived temperature reconstructions. Analysis of the global energy budget during the last millennium indicates that Little Ice Age (LIA) cooling is largely driven by volcanic forcing (comprising an average of 82%, 54% and 77% of the total forcing among models for the 1200-1450 CE, 1400-1650 CE and 1600-1850 CE periods, respectively, relative to 950-1200 CE), while contributions due to changes in total solar irradiance (15%, 30% and 10%) and greenhouse gas concentrations (3%, 16% and 13%) are substantially lower. Focusing on the period from 1600-1850 CE, the combination of these forcings directly contributes to 40%, on average, of the global cooling during the LIA, while the remainder of the cooling arises due to the Planck response and the sum of the climate feedbacks. Dominant positive feedbacks include the water
vapor feedback, which contributes 29% of the global cooling and arises due to decreased water vapor concentration associated with the vertically-integrated cooling and the surface albedo feedback, which contributes 14% of the global cooling and arises due to high latitude sea ice expansion and increased snow cover. Lapse rate changes contribute an additional 7% of the global cooling; the positive lapse rate feedback in the last millennium simulations is in contrast to the negative lapse rate feedback that is observed in the CMIP5 future climate simulations and arises due to greater cooling near the surface than aloft throughout the mid- and high-latitudes.

**Introduction**

The temporal evolution and spatial structure of temperature variability over the past millennium serve as important constraints for separating anthropogenic impacts from natural climate variability. Reconstructions point to two major climate epochs during the preindustrial era— the Medieval Climate Anomaly (MCA; ca. 900-1250 CE) and the Little Ice Age (LIA; ca. 1300-1850 CE; e.g. Crowley, 2000; Jones et al., 1998; Lamb, 1965; Mann et al., 2009). Paleoclimate proxy records demonstrate substantial heterogeneity in the timing, amplitude, and spatial extent of the MCA and LIA, thus these periods are characterized by regionally specific temperature departures from an overall global cooling trend over the last millennium (PAGES 2k
Consortium, 2013). Temperature reconstructions across the globe generally indicate colder conditions beginning ca. 1200-1580 CE. In the North Atlantic region, colder air and sea surface temperatures (Fig. 1B-D; Cunningham et al., 2013; Larsen et al., 2011; Stuiver et al., 1997) and increased sea ice cover (Fig. 1A; Masse et al., 2008; Ogilvie and Jonsson, 2001) appear to have begun ca. 1200 CE. Over continental regions, cooling in North America and the Southern Hemisphere lagged that in the Arctic, Europe and Asia (Fig. 1; PAGES 2k Consortium, 2013). Though interspersed with periods of warmth, cold conditions appear to have prevailed until the late 19th century.

Solar, volcanic and orbital forcings combined with atmosphere, ocean and sea-ice feedbacks are thought to have driven large scale cooling during the LIA (Bond et al., 1997; Briffa et al., 1998; Bianchi and McCave, 1999; Broecker, 2000; Kaufman et al., 2009; Palastanga et al., 2009; Zhong et al., 2010; Miller et al., 2012; Wanamaker et al., 2012; Lehner et al., 2013; PAGES 2k Consortium, 2013; Schleussner and Feulner, 2013). In addition, anthropogenic forcing (Schurer et al., 2013) and internal variability (Crowley, 2000; Wanner et al., 2011) may have played important roles in contributing to this cold period. For instance, while the global signal of the LIA has been attributed to the role of external forcings, the spatial and temporal heterogeneity of climate change during this period is likely a result of a complex array of forcing, feedbacks and internal variability in the climate system (Kaufman et al., 2009; Lehner et al., 2013).
Recent improvements in proxy-derived constraints of climate forcings over the last millennium as well as an increasing number of transient climate model simulations with last millennium forcings have enabled new approaches to be used to examine the causes of the LIA. Implementation of transient model simulations with individual forcings enable quantification of the relative contribution of individual forcings, while statistical techniques enable estimation of the role of internal variability versus external forcing (Schurer et al., 2013). However, to date, only a limited number of such single-forcing model simulations have been performed. Because quantification of the driving mechanisms of the LIA remains elusive, the mechanisms responsible for shaping the global characteristics of the LIA have yet to be determined.

Here we quantify the role of radiative forcings and climate feedbacks to global cooling over the last millennium in transient climate model simulations from the CMIP5/PMIP3 archive. The contribution of the forcing and feedback mechanisms during the LIA is quantified through the use of the Approximate Partial Radiative Perturbation (APRP) method and radiative kernels. We compare the global temperature anomalies in the last millennium simulations to those in the pre-industrial control runs of the same models to assess the role of internal variability in the climate system.

2. Methods

Model simulations and forcings
In this analysis, we use output from transient climate simulations of the last millennium and unforced 850-1000 year-long control simulations with preindustrial boundary conditions from seven different Atmosphere-Ocean General Circulation Models (AOGCMs) from the CMIP5/PMIP3 (Coupled Model Intercomparison Project Phase 5/Paleoclimate Modeling Intercomparison Project Phase 3). These models are: CCSM4 (Landrum et al., 2013), GISS-E2-R forcing ensemble members r1i1p121 and r1i1p124 (GISS121 and GISS124 hereafter), MPI-ESM-P (MPI hereafter), IPSL-CM5A-LR (IPSL hereafter; Dufresne et al., 2013), CSIRO-Mk3L-1-2 (CSIRO hereafter), and HadCM3. These models were chosen based on the availability of data at the time of analysis. MIROC-ESM and FGOALS-s2 were omitted due to long-term drifts in global mean surface air temperature in their preindustrial control simulations. Details of the models, external forcing and references can be found in Table 1.

Different forcing data sets were imposed in the last millennium simulations of different GCMs, following the protocols of PMIP3 (https://pmip3.lsce.ipsl.fr/wiki/doku.php/pmip3:design:lm:final) as discussed by Schmidt et al. (2012) and outlined in Table 1. Also shown in Figure 2 are the forcings (from top to bottom): volcanic aerosols, greenhouse gas (CH$_4$, CO$_2$, and N$_2$O) concentrations, solar variability associated with changes in orbital configuration and solar output, and (not shown) anthropogenic land-use changes.
a. Volcanic forcing

Reconstructions of volcanic aerosols used in the last millennium simulations (Fig. 2A) are derived from either the Gao et al. (2008; hereafter GRA) or Crowley et al. (2000; hereafter CEA) data sets of sulfate loading and AOD, respectively, as indicated in Table 1 (Schmidt et al., 2011; https://pmip3.lsce.ipsl.fr/). The CEA data set is based on 13 Greenland and Antarctic ice cores. AOD and effective radius are given in 10-day intervals in four equal area latitude bands. The GRA data set is based on 54 ice cores, 32 from Arctic and 22 from Antarctica. Sulfate loading is provided in the GRA data set as a function of month, latitude in 10 bands, and height from 9 to 30 km at 0.5 km resolution. Four models (GISS 121, GISS 124, MPI, HadCM3) prescribed volcanic aerosols in terms of AOD and aerosol effective radius from CEA. One model directly prescribed sulfate loading from GRA (CCSM4). IPSL also prescribed sulfate loading but an error in the implementation resulted in different loading than that in the GRA data set (J.-L. Dufresne, personal communication). One model (CSIRO) estimated globally averaged forcing from GRA sulfate loading and converted this quantity to a total solar irradiance (TSI) anomaly (Phipps et al., 2012).

b. Trace gas forcing

Changes in concentration of the principle well mixed GHGs (CO$_2$, CH$_4$ and N$_2$O) over the last millennium (Fig. 2D) are related to both natural
variations as well as anthropogenic factors in the latter part of the last millennium (Gerber et al., 2003). Reconstructions are derived from high-resolution ice cores in Antarctica and smoothed to retain only decadal-scale and longer variations (Joos and Spahni, 2008; Schmidt et al., 2011). All models use the same set of GHG concentrations, as described in Schmidt et al. (2011). In addition, one model (MPI) parameterizes ozone variations as a function of changes in solar irradiance based on the results of Shindell et al. (2006).

c. Orbital forcing

Orbital forcing during the last millennium was primarily characterized by changes in precession. From 850 CE to the present, a ca. 20-day shift in perihelion (from Dec. 15th to Jan. 4th) occurred, leading to an increase in insolation in early NH summer relative to the late NH summer (Schmidt et al., 2011). Small decreases in eccentricity and obliquity also occurred (Schmidt et al., 2011). These changes gave rise to a ca. 3 W/m² decrease in insolation in boreal summer (JJA) through the last millennium (Fig. 2E).

d. Solar forcing

Changes in TSI are prescribed using either the Vieira et al (2011; VSK) or Steinhilber et al. (2009; SBF) reconstruction as described in Schmidt et al. (2011). In some models, “background” variations of TSI (variations not tied to
the solar cycle) are taken from Wang et al. (2005; WLS), as indicated in Table 1. TSI anomalies are shown in Fig. 2F.

e. Land use forcing

Reconstructions of land use and land cover are available for the last three centuries based on published maps of agricultural areas and for earlier periods based on scaling agricultural activity with population on a per country basis (Pongratz et al., 2008; hereafter referred to as PEA). The resultant data set provides annual maps of cropland, C3 pasture and C4 pasture, which influence the surface albedo, water cycle, surface roughness and soil characteristics (Schmidt et al., 2011).

Description of APRP method

The Approximate Partial Radiative Perturbation (APRP) method decomposes changes in top of the atmosphere (TOA) shortwave (SW) energy fluxes into individual radiative forcing terms and climate feedback terms (Taylor et al., 2007). It is based on a simple shortwave radiative model of the atmosphere, in which the influence of changes in surface albedo, shortwave absorption and scattering on the top of the atmosphere energy budget are diagnosed at every grid cell from all-sky and clear-sky GCM output. APRP analysis relies on the use of a single layer radiative transfer model that is tuned to mimic the radiation code of the GCM. Specifically, three parameters
in the single layer model are calculated (surface albedo, atmospheric scattering coefficient and atmospheric absorption coefficient) to ensure that the surface and top of the atmosphere shortwave fluxes are consistent with those in the GCM. These single layer model parameters are calculated for two time periods, representing the control and perturbed conditions. These parameters are then individually perturbed in the single layer model by the amount they change between the control and perturbed periods of the GCM simulation and the influence of these changes on the TOA shortwave flux in the single layer model is calculated. In this way, the single layer model enables the effects of changes in surface albedo, atmospheric absorption, atmospheric scattering and clouds to be isolated from one another using a simple and efficient method. This method is similar to the Partial Radiative Perturbation (PRP) method, but whereas the PRP method requires running the GCM’s radiation code offline with the various radiation properties individually perturbed, APRP calculations are far less computationally expensive and require far less data from the full GCM simulations (only monthly clear-sky and full-sky radiative flux fields at the surface and TOA are needed). A comparison between full PRP and APRP analyses of global warming and Last Glacial Maximum simulations with two GCMs demonstrated that the differences between them were typically only a few percent (Taylor et al., 2007). Further details of the APRP method can be found in Taylor et al. (2007).
Calculation of feedbacks using APRP and radiative kernel method

Quantification of the forcing and feedbacks in the CMIP5/PMIP3 last millennium simulations can be estimated using the Approximate Partial Radiative Perturbation (APRP) method to decompose changes in the global shortwave (SW) TOA energy budget (Taylor et al., 2007) and using the radiative kernel method to decompose changes in the global longwave (LW) TOA energy budget (Soden and Held, 2006; Soden et al., 2008; Shell et al., 2008).

The response of the TOA global energy budget to a radiative forcing can be described as follows: the climate system responds to a radiative imbalance through a change in global mean surface temperature, which produces a climate response that opposes the forcing. The climate response is represented as a linear change in global mean surface temperature (ΔT_s) with the multiplicative constant given by the climate feedback parameter (λ). This is described by the following equation:

\[
\Delta R_{\text{imb}} = \Delta R_{\text{forc}} + \Delta R_{\text{resp}} = \Delta R_{\text{forc}} + \lambda \Delta T_s
\]
where $\Delta R_{\text{forc}}$ represents the radiative forcing, $\Delta R_{\text{resp}}$ represents the TOA energy fluxes due to climate feedbacks (i.e. the climate response), and $\Delta R_{\text{imbal}}$ represents the remaining imbalance in the Earth’s TOA energy budget.

We define the total global-mean, annual-mean net TOA forcing as the sum of the solar, volcanic, and GHG forcings ($\Delta R_{\text{forc}} = \Delta R_{\text{solar}} + \Delta R_{\text{volc}} + \Delta R_{\text{GHG}}$). (Given the magnitude and distribution of the surface albedo changes in comparison to land cover changes (Pongratz et al., 2008), we assume that global-mean forcing from land use changes is small.) Solar forcing is the net change in SW$_{\text{TOA}}$ due to changes in insolation:

$$\Delta R_{\text{solar}} = (1-A)*\Delta S , \tag{2}$$

where $S =$ insolation (in W/m$^2$) and $A =$ planetary albedo. This solar forcing term is further decomposed into changes in orbital configuration and solar output using equations from Berger et al. (1978). The GHG forcing is estimated using the formulas from Myhre et al. (1998). The volcanic forcing is estimated using the method of APRP as described below.

The APRP method decomposes the changes in net SW$_{\text{TOA}}$ into components associated with changes in insolation and changes in planetary albedo ($\Delta A$) and further decomposes changes in planetary into contributions from cloud and non-cloud scattering and absorption, surface albedo, and non-linear (residual) effects:
\[ \Delta R_{SW,TOA} = \Delta [S(1-A)] \]  
\[ \Delta R_{SW,TOA} \approx \Delta S(1-A) - S\Delta A \]
\[ \approx \Delta R_{\text{Solar}} + \Delta R_{\text{Plan Alb}} \]
\[ \Delta R_{\text{Plan Alb}} = S \ast (\Delta A_{SW,\text{cloud}} + \Delta A_{SW \text{ noncloud scat}} + \Delta A_{SW \text{ noncloud abs}} + \Delta A_{\text{sfc alb}} + \Delta A_{SW,\text{res}}) \]

where \( \Delta R_x \) indicates the contribution of \( x \) to TOA energy flux changes and \( \Delta A_x \) indicates the contribution of \( x \) to planetary albedo changes. Equations for the \( \Delta A_x \) terms can be found in Taylor et al. (2007). Atmospheric non-cloud scattering is driven primarily by changes in atmospheric aerosols, thus the APRP term that describes changes in net SW\_TOA due to non-cloud scattering can be attributed to changes in volcanic aerosol injection during the last millennium. In this way, the non-cloud scattering term from the APRP analysis can be used to quantify volcanic forcing (\( \Delta R_{\text{volc}} = \Delta R_{SW \text{ non-cloud scat}} \)).

The climate feedback parameter can also be split into various contributions by neglecting interactions among feedback variables. Here we decompose the total effective climate feedback parameter into three SW components (surface albedo feedback, SW non-cloud absorption and SW cloud feedback) using the above APRP analysis and four LW components (Planck response, lapse rate feedback, LW water vapor feedback and LW cloud feedback) using the radiative kernel technique (Soden and Held, 2006; Soden
Atmospheric non-cloud absorption is primarily driven by changes in the absorption of incoming SW radiation by atmospheric water vapor, so the APRP non-cloud absorption term is therefore interpreted as the SW water vapor feedback (implicit in this assumption is that the contribution of SW absorption by volcanic aerosols is small relative to water vapor changes). SW and LW residual terms are also calculated, which account for nonlinearities in the climate feedbacks:

\[ \Delta R_{\text{resp}} = \lambda \Delta T_s = \Delta T_s \ast (\lambda_{\text{SW}} + \lambda_{\text{LW}}) \]  

(4)

\[ \lambda_{\text{SW}} = \lambda_{\text{sfc alb}} + \lambda_{\text{SW,q}} + \lambda_{\text{SW,cloud}} + \lambda_{\text{SW,res}} \]  

(5)

\[ \lambda_{\text{LW}} = \lambda_{\text{Planck}} + \lambda_{\text{lapse rate}} + \lambda_{\text{LW,q}} + \lambda_{\text{LW,cloud}} + \lambda_{\text{LW,res}} \]  

(6)

where q is specific humidity. The feedbacks are estimated as follows:

\[ \lambda_x = \frac{\Delta R_x}{\Delta T_{as}} \]  

(7)

where the SW feedbacks \( \Delta R_x \) (\( \Delta R_{\text{sfc alb}} \), \( \Delta R_{\text{SW,q}} \), \( \Delta R_{\text{SW,cloud}} \) and \( \Delta R_{\text{SW,residual}} \)) are obtained from APRP output and \( \Delta T_{as} \) is the global-mean, annual averaged surface air temperature change between two time periods.

The global LW energy budget is evaluated using the radiative kernel technique of Soden and Held (2008). Under this framework, climate feedbacks are represented as the product of two terms: the first is the radiative kernel, which is a weighting term that describes the TOA flux perturbation due a small change in a particular climate variable (e.g. specific humidity), the
second is the change in that climate variable in the full GCM simulation. The kernels used in this analysis are from Karen Shell (2008) and were calculated by perturbing the climate variables in CAM3 by a small amount and measuring the TOA flux response. Studies have shown that radiative kernels calculated from different models are highly similar (Soden et al., 2008; Vial et al., 2013).

The LW feedbacks are estimated using the radiative kernel method as follows:

\[
\Delta R_{\text{Planck}} = \frac{\partial R}{\partial T} \Delta T_s = (K_{Ta} + K_{Ts}) \Delta T_s \quad (8)
\]

\[
\Delta R_{\text{lapse rate}} = \frac{\partial R}{\partial T} (\Delta T_a - \Delta T_s) = K_{Ta} (\Delta T_a - \Delta T_s) \quad (9)
\]

\[
\Delta R_{LW,q} = \frac{\partial R}{\partial \ln(q)} \Delta \ln(q) = K_q \Delta \ln(q) \quad (10)
\]

where \(K_{Ta}(mo, x, y, z), K_q(mo, x, y, z)\) and \(K_{Ts}(mo, x, y)\) are the all-sky kernels, \(\Delta T_s(mo, x, y)\) is the monthly climatology of skin temperature and \(\Delta T_a(mo, x, y, z)\) is the air temperature and \(q(mo, x, y, z)\) is the specific humidity. To calculate the Planck response, \(\Delta T_s\) is applied to all vertical levels in the atmosphere.

Cloud feedbacks cannot be calculated directly from radiative kernels because of strong nonlinearities that arise from cloud masking. Following Soden et al. (2008), we adjust the cloud radiative effect by correcting for non-
cloud feedbacks. The clear-sky GHG forcing is calculated following Soden et al. (2008): $\Delta R_{\text{GHG}}^{CS} = 1.16 \times \Delta R_{\text{GHG}}$. The LW residual term is calculated as the difference between the net LW response ($\Delta R_{\text{LW,resp}} = \Delta R_{\text{LW,imbal}} - \Delta R_{\text{GHG}}$) and the sum of the LW climate feedbacks:

$$
\lambda_{\text{LW, res}} = [\Delta R_{\text{LW,imbal}} - \Delta R_{\text{GHG}} - \overline{\Delta T_{as}^*} (\lambda_{\text{Planck}} + \lambda_{\text{lapse rate}} + \lambda_{\text{LW,q}} + \lambda_{\text{LW,cloud}})] / \overline{\Delta T_{as}} . \quad (11)
$$

Finally, global cooling contributions due to radiative forcings, changes in ocean and atmospheric heat uptake and climate feedbacks are calculated (following Vial et al., 2013 and Fedl and Roe, 2013):

$$
\overline{\Delta T_{as}} = [\Delta R_{\text{imbal}}/ \lambda_{\text{Planck}} - \Delta R_f/ \lambda_{\text{Planck}} - \sum_{\lambda \neq \text{Planck}} \frac{\lambda}{\lambda_{\text{Planck}}} \overline{\Delta T_{as}}^*] \quad (12)
$$

$$
\Delta T_{as} = \overline{\Delta T_{\text{atm+ocn uptake}}} + \overline{\Delta T_{\text{forc}}} + \overline{\Delta T_{\text{fdbs}}} .
$$

The fraction of global cooling due to individual forcings and feedbacks is calculated as:

$$
f_{\overline{\Delta T_{as}}} (%) = \frac{\overline{\Delta T_x}}{\sum \overline{\Delta T_x}} \times 100 , \quad (13)
$$

where $\overline{\Delta T_x}$ includes only the negative (cooling) terms.

3. Results/Discussion

3.1 CMIP5 Temperature Trends Over the Last Millennium
Fig. 2C shows the time series of global mean surface temperature anomaly in the last millennium and preindustrial control simulations, (annually averaged data were smoothed with a Gaussian filter; $\sigma = 3$ years) Relative to the period from 850-1200 CE, all CMIP5/PMIP3 simulations generally demonstrate colder global temperatures ca. 1250 – 1850 CE. The period from 1200-1850 CE is characterized by substantial interannual- to centennial-scale temperature variability with pronounced minima ~1230-1290, 1450, 1600-1700, and 1820-1840 CE, corresponding with volcanically active periods. Compared to the preindustrial control simulations from the same models, the last millennium simulations demonstrate global temperature anomalies outside the range of natural variability during volcanically active periods. In addition, remarkably good agreement is generally demonstrated between modeled and proxy reconstructed NH temperature anomalies over the last millennium, particularly with regard to the low-frequency variations (Fig. 3; Frank et al., 2010). That temperature anomalies associated with large volcanic events tend to be substantially larger in the models than in the reconstructions may be related to issues of tree growth subsequent to large volcanic events affecting tree ring-based temperature reconstructions (Mann et al., 2012) and/or the models’ overestimation of volcanic forcing during large volcanic events due to treatment of volcanic aerosols (Timmreck et al., 2009) and uncertainties in the volcanic reconstructions (Sigl et al., 2014).
To quantify the contribution of radiative forcings and climate feedbacks to global cooling over the last millennium, we compare the energy budget during the LIA to the MCA, initially defining the LIA as the period from 1600-1850 CE and later extending our analysis of the LIA to the time periods from 1400-1650 CE and 1200-1450 CE. For all cases we define the MCA as the period from 950-1200 CE.

While the 1600-1850 CE period is characterized by centennial-scale global temperature minima in all CMIP5/PMIP3 models, the amplitude and spatial pattern of surface temperature anomalies (relative to the MCA) displays marked differences among models (Fig. 4). The regional temperature differences are reflected in changes in sea ice concentration (Fig. 5); in the Arctic, sea ice concentration increases in all models, whereas in the Antarctic, sea ice concentration increases in some models (CCSM4, CSIRO, HadCM3) and decreases in others (GISS121 and GISS124). Sources of these inter-model differences are discussed in Section 3.3.

3.2 Attribution of LIA climate change in CMIP5/PMIP3 models

Climate forcing during the last millennium is comprised of contributions from changes in solar output, orbital configuration, stratospheric sulfate aerosols associated with large volcanic eruptions, changes in trace gases and changes in land use through its impact on surface albedo and evapotranspiration (Table 1; Bony et al., 2006; Schmidt et al.,
2012). Given the magnitude and distribution of the surface albedo changes, we assume that the influence of land use changes is small. The global cooling contributions from each of these forcings is shown in Fig. 7 while the global TOA energy fluxes are shown in Fig. S1.

A TOA energy budget analysis of the CMIP5/PMIP3 last millennium simulations indicates that volcanic forcing was the primary driver of LIA cooling in the models. During the 1600-1850 CE period, volcanic forcing accounted for 77% of the total forcing on average (ranging from 50-84% or \(-0.07 \text{ to } -0.32 \text{ Wm}^{-2}\), across models; Fig. 6). This forcing directly contributed to 31% (on average) of the global cooling during the LIA (Fig. 7; Table 2). In comparison, solar and GHG forcing were substantially weaker, comprising an average of 10% and 13% of the total forcing, respectively. Solar and GHG forcing contributed on average, 4% and 5%, respectively, of the global cooling during the period from 1600-1850 CE. Globally averaged solar forcing is driven by changes in solar output over the last millennium as changes in orbital parameters impart insignificant forcing (Fig. 6). However it is important to note that orbital forcing likely gives rise to global feedbacks. For instance, precessional forcing imposes a ca. 3 Wm\(^{-2}\) decrease in insolation in boreal summer (JJA) at 65°N through the last millennium (Fig. 2E) which likely contributed to sea ice growth and colder temperatures in the Arctic (Kaufman et al., 2009).
A number of positive climate feedbacks reinforce the radiative forcing during the LIA. The largest positive feedback is the LW water vapor feedback, which is responsible for 20%, on average, of the global cooling (Fig. 7; Table 2). The LW water vapor feedback occurs in response to the decreased atmospheric water vapor concentration; the saturation vapor pressure decreases as the atmosphere cools as given by the Clausius-Clapeyron equation. In addition, the SW water vapor feedback represents a lesser but globally important positive feedback in all of the models (contributing 9%, on average, to the global cooling), consistent with decreased absorption of incoming SW radiation by atmospheric water vapor during the LIA.

In global warming simulations, the positive water vapor feedback is due to both the vertically uniform atmospheric warming as well as the vertical redistribution of water vapor (as robust decreases in tropical lapse rate from enhanced warming aloft lead to an upward shift in the water vapor distribution). The large positive water vapor feedback (1.71 ±0.13 Wm⁻²K⁻¹; Vial et al., 2013), is partially offset by a negative lapse rate feedback, leading to a net feedback of 1.05 ±0.06 Wm⁻²K⁻¹ in the CMIP5 4xCO₂ simulations (Vial et al., 2013). In contrast, the CMIP5/PMIP3 last millennium simulations demonstrate lapse rate and water vapor feedbacks that reinforce one another. The positive lapse rate feedback arises due to greater cooling near the surface than aloft poleward of ca. 30-40° latitude (Fig. S2) and contributes 7% on average to the global cooling (Fig. 7; Table 2). Combined, the water vapor and
lapse rate feedbacks are responsible for 36%, on average, of the global cooling during the LIA. Thus, while the total (SW + LW) water vapor feedback (1.25 ±0.19 Wm⁻²K⁻¹; Fig. 8) is substantially less than that reported from the CMIP5 4×CO₂ simulations, the combined water vapor and lapse rate feedback is 2.32±0.34 Wm⁻²K⁻¹ in the last millennium runs, more than twice that obtained from global warming simulations.

Second to the LW water vapor feedback, the next largest global feedback is the surface albedo feedback, which is responsible for 14% of the global cooling on average (Fig. 7; Table 2). The surface albedo feedback arises primarily from increases in high latitude sea ice during the LIA in CCSM4 and MPI. In HadCM3, GISS121 and GISS124 this feedback is equally or predominately due to increased surface albedo (presumably from increased snow cover) over Eurasia and North America (data not shown). In IPSL, the SW cloud feedback is the dominant positive SW feedback (Fig. 7).

Not unexpectedly, clouds have a varied response among models, most notably in the SW cloud feedbacks. LW cloud feedbacks are positive in all models, consistent with the CMIP5 4×CO₂ simulations (Vial et al., 2013) and likely due to lower cloud tops in the colder LIA atmosphere. The LW cloud feedback contributes 6%, on average, to the global cooling. In contrast, in all models aside from IPSL, the SW cloud feedback is a negative feedback, generally arising from decreased cloud fraction in the tropics and high latitudes. The positive SW cloud feedback in IPSL is the dominant positive
feedback and arises from increased cloud fraction in the midlatitudes, particularly in the SH (data not shown).

The SW residual term ranges from -0.01 – -0.02 Wm⁻²K⁻¹, or 2 – 3% of the total SW feedback parameter while the LW residual term amounts to -0.04 – -0.64 Wm⁻²K⁻¹, or 2 – 25% of the total LW feedback parameter. The LW residual is a consequence of the linear framework of the feedback analysis and may additionally be biased by LW absorption by volcanic aerosols, which has been neglected in this analysis. However, our average residual (16%) is smaller than that reported from the CMIP5 4xCO2 simulations (23%; Vial et al., 2013).

3.3 Sources of inter-model differences in LIA climate change

While the 1600-1850 CE period is characterized by global cooling relative to the MCA in all models, the amplitude of the cooling differs by more than a factor of 2 among models (Fig. 7). Factors that may be responsible for the differences in the LIA global mean temperature change among models include differing climate forcings, differing climate feedbacks and differing efficiencies with which they transfer heat into the ocean. However, ocean heat uptake during the LIA is small in all the models, as demonstrated by small changes in surface energy fluxes (Fig. S1). Thus, the TOA radiative forcings and feedbacks are primarily responsible for the inter-model spread in global cooling during the LIA. Among radiative forcings, differences in volcanic
forcing drive the inter-model spread in total forcing (Fig. 6). Differences in volcanic forcing across models could arise from differences in the volcanic data set used (i.e. uncertainty in volcanic aerosol reconstructions) as well as from differences in the treatment of volcanic aerosols in the models. The CEA data set of AOD and effective radius indicates substantially lower AOD anomalies during large volcanic events (Fig. 2A) but a greater increase in the number of volcanic events during the LIA relative to the MCA. The use of different volcanic data sets is responsible for a substantial portion of the differences in volcanic forcing among models (e.g. compare the volcanic forcing in CCSM4 to GISS/MPI/HadCM3 in Fig. 6). However, an even larger contribution to the inter-model spread in volcanic forcing arises from the differing treatment of volcanic aerosols among models that use the same volcanic reconstruction. For instance, volcanic forcing in CSIRO (which imposes volcanic forcing as a globally-averaged perturbation to TSI) is up to 50% lower than that in CCSM4 (which has an atmospheric chemistry scheme that models the transport, processing and radiative properties of the volcanic aerosols; Fig. 6).

In addition to differences in radiative forcing among models, differences in the strength of climate feedbacks add to the inter-model spread in LIA cooling. The total effective feedback parameter differs by a factor of 2 across models (ranging from -1.03 to -2.12 Wm$^{-2}$K$^{-1}$; Fig. 8). The Planck response varies little among models, as expected (-3.41 ±0.05 Wm$^{-2}$K$^{-1}$). The largest spread in climate feedbacks among models occurs in the SW cloud feedback (-
0.20 ±0.47 Wm$^{-2}$K$^{-1}$) and the lapse rate feedback (0.36 ±0.25 Wm$^{-2}$K$^{-1}$). The large spread in the SW cloud feedback is a common feature in GCM simulations (Crucifix et al., 2006; Soden and Held, 2006; Vial et al., 2013). The large spread in the lapse rate feedback is likely in part due to the varied response of high latitude sea ice among models (Fig. 5; Fig. S2).

3.4 Sensitivity of results to definition of LIA

Attribution of global cooling during the period from 1200-1450 CE and 1400-1650 CE is similar to that for the 1600-1850 CE period. During all three periods, volcanic forcing is the dominant forcing, with a multi-model mean of 81%, 55% and 77% of the total forcing during the 1200-1450 CE, 1400-1650 CE and 1600-1850 CE periods, respectively. The 1600-1850 CE period is characterized by the largest total forcing, on average (-0.32 W/m$^2$). While the 1400-1650 CE period is characterized by the weakest total forcing (-0.19 W/m$^2$), but the strongest relative contribution from solar forcing (30%; Fig. 6). Total greenhouse gas forcing is similar between 1600-1850 CE and 1400-1650 CE (-0.03 Wm$^{-2}$) but slightly weaker in 1200-1450 CE (-0.01 Wm$^{-2}$).

In addition to the total forcing varying substantially among the different LIA epochs, the mean effective climate feedback parameter also varies across epochs (between -1.42 – -2.16 Wm$^{-2}$K$^{-1}$). Differences in the effective climate sensitivity arise primarily due to changes in the SW cloud feedback and the net LW feedback. Ocean heat uptake is small across all three
periods, as demonstrated by changes in global mean surface energy fluxes of less than $\pm 0.05 \text{ Wm}^{-2}$ (data not shown).

5. Conclusions

The CMIP5/PMIP3 last millennium simulations demonstrate global cooling during the period ~ 1200-1850 CE that generally agrees well with proxy based temperature reconstructions. Analysis of the global TOA energy budget during the last millennium using APRP and radiative kernel techniques indicates that the LIA cooling in the CMIP5/PMIP3 models is largely driven by volcanic forcing, as volcanic forcing contributes 82%, 54% and 77% of the total forcing in the multi-model mean for the 1200-1450 CE, 1400-1650 CE and 1600-1850 CE periods, respectively (relative to 950-1200 CE). Relative contributions from solar (15%, 30% and 10%) and greenhouse gas (3%, 16% and 13%) forcing are substantially smaller.

A feedback analysis of the 1600-1850 CE period demonstrated that the dominant climate feedbacks that reinforce the global cooling include the water vapor and lapse rate feedbacks, which combined are responsible for 36% of the LIA cooling in the models. The positive LW and SW water vapor feedbacks are a consequence of the decrease in water vapor concentration associated with the vertically-integrated cooling while the positive lapse rate feedback arises due to greater cooling near the surface than aloft poleward of ca. 30-40° latitude. The dominant positive SW feedback is the surface albedo feedback,
which is responsible for 14% of the global cooling on average and arises from sea ice growth and increased snow cover during the LIA. Increased sea ice and snow cover may be due to both globally-averaged forcing (i.e. volcanic forcing) as well as precessional forcing (through decreased NH high latitude insolation in boreal summer). LW cloud feedbacks provide an additional 6% of the global cooling on average and are consistent with lower cloud tops in the colder LIA atmosphere.

There are several points to consider when interpreting the results of the CMIP5/PMIP3 last millennium simulations in light of the paleoclimate record. Firstly, large uncertainties exist in the characterization of volcanic forcing that may not be adequately represented in the last millennium simulations. Large uncertainties exist in reconstructions of sulfate concentration, AOD and aerosol effective radius as a function of time, latitude and height, all of which exert important controls on the climate system. These uncertainties are compounded by the CMIP5 models’ poor representation of the dynamical response of the atmosphere to volcanic eruptions (Driscoll et al., 2012). Further, large uncertainties remain concerning the magnitude of changes in insolation over the last millennium. In particular, some TSI reconstructions suggest that changes in insolation during the LIA were an order of magnitude larger than those used in the CMIP5/PMIP3 last millennium simulations (Shapiro et al., 2011). This uncertainty, in conjunction with the fact that only one of the CMIP5/PMIP3 last millennium runs included solar-driven ozone
variations (Shindell et al., 2006) leave open the possibility that solar forcing may have played a larger role in LIA cooling than suggested by these model simulations.

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References

Fig. 1: Proxy records of climate change over the past millennium: A) record of Arctic sea ice changes based on the relative abundance of IP25, a biomarker produced by sea ice algae, in a sediment core from the north coast of Iceland (Massé et al., 2008); B) record of ice cap extent in central Iceland from Hvitárvatn Lake sediment varve thickness (Larsen et al., 2011); C) multi-proxy NE Atlantic SST composite (Cunningham et al., 2013); D) δ18Oice from the GISP2 ice core (Stuiver et al., 1997), a proxy for surface air temperature in Greenland; E) NH temperature anomalies based on a multi-proxy network and the climate field reconstruction method (Mann et al., 2009).
Fig. 2: Climate forcings and global surface air temperature through the last millennium in the CMIP5 models: A) Aerosol optical depth from the Crowley et al. (2000) data set (blue bars) and estimated from the Gao et al. (2008) data set (red bars) by dividing sulfate loading by 150 Tg (Stothers, 1984). B) Globally averaged surface air temperature anomaly (relative to 950-1200 CE) in each last millennium simulation. C) Globally averaged surface air temperature anomaly for the multi-model ensemble mean of the last millennium simulations (blue) and the control simulations (green) where solid lines represent the multi-model mean and shading represents ±1σ. Temperature anomalies are calculated relative to years 100-350 in each data set and annually averaged data were smoothed with a Gaussian filter with σ = 3 years. D) Concentration of CO₂, CH₄, and N₂O. E) Changes in insolation at 65°N in JJA (blue) and 65°S in DJF (red). F) Globally averaged insolation anomalies (relative to 950-1200 CE) for the solar forcing data sets outlined in Table 1.
Fig. 3. Mean NH temp anomaly from CMIP5/PMIP3 models (blue) and from 521 ensemble temperature reconstructions (black), based on 9 different large-scale NH temperature reconstructions spanning the past millennium (Frank et al., 2010). Solid lines represent the multi-model/multi-reconstruction mean and shading represents ±1σ.
Fig. 4. LIA (1600 – 1850 CE) minus MCA (950-1200 CE) surface air temperature changes in the CMIP5/PMIP3 simulations.
Fig. 5. LIA (1600 – 1850 CE) minus MCA (950-1200 CE) sea ice concentration changes in the CMIP5/PMIP3 simulations.

Fig. 6. Decomposition of radiative forcings in the CMIP5/PMIP3 last millennium simulations into contributions from volcanic aerosols, solar output, orbital configuration and the well-mixed GHGs for the period from 1200-1450 CE (P1), 1400-1650 CE (P2) and 1600-1850 CE (P3), relative to the MCA (950-1200 CE). Percentages in the brown, yellow and blue boxes represent fractional contributions to total forcing from volcanic, total solar (orbital and TSI) and total GHG (CO₂, CH₄ and N₂O) forcings, respectively. Values below the bars represent total forcing in W/m².
Fig. 7. Global cooling contributions from Eqn. 12 due to volcanic, solar and GHG forcings and the SW and LW feedbacks compared to the total globally averaged temperature change between the LIA (1600 – 1850 CE) and MCA (950-1200 CE). The difference between the total cooling and the sum of the forcings and feedbacks is the Planck response.

Fig. 8. LIA climate feedback parameters calculated from the CMIP5/PMIP3 simulations for the SW (surface albedo, SW cloud, SW absorption, SW residual), LW (total LW, Planck response, lapse rate, LW water vapor and LW cloud, LW residual) and total feedbacks.
Supplementary Material

Fig. S1. LIA (1600 – 18050AD) minus MCA (950-1200 AD) globally averaged TOA energy flux changes ($\Delta R$) due to forcings (volcanic, solar, greenhouse gas), SW feedbacks (surface albedo, SW cloud, SW absorption, SW residual) and LW feedbacks (Planck, lapse rate, LW water vapor, LW cloud, LW residual) and globally averaged surface energy flux ($\Delta E_{sfc}$).

Fig. S2. LIA (1600 – 1850AD) minus MCA (950-1200 AD) changes in tropospheric lapse rate (calculated as a pressure-weighted average between 1000-200 mbar).