Climate Variability and Change since 850 C.E.: An Ensemble Approach with the Community Earth System Model (CESM)

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Abstract

The climate of the past millennium provides a baseline for understanding the background of natural climate variability upon which current anthropogenic changes are superimposed. As this period also contains high data density from proxy sources (e.g. ice cores, stalagmites, corals, tree rings, and sediments), it provides a unique opportunity for understanding not only global but regional-scale climate responses to natural forcing. Towards that end, an ensemble of simulations with the Community Earth System Model (CESM) for the period 850-2005 (the CESM Last Millennium Ensemble, or CESM-LME) is now available to the community. This ensemble includes simulations forced with the transient evolution of solar intensity, volcanic emissions, greenhouse gases, aerosols, land use conditions, and orbital parameters, both together and individually. The CESM-LME thus allows for evaluation of the relative contributions of external forcing and internal variability to changes evident in the paleoclimate data record, as well as providing a longer-term perspective for understanding events in the modern record. It also constitutes a dynamically consistent framework within which to diagnose mechanisms of regional variability. Results demonstrate an important influence of internal variability on regional responses of the climate system during the past millennium. All the forcings, and particularly that associated with large volcanic eruptions, are found to be influential during the preindustrial period, while anthropogenic greenhouse gas and aerosol changes dominate the forced variability of the mid to late 20th century.
Capsule

The CESM-LME community-modeling project gives the research community a resource for better understanding both proxy records and climate variability and change since 850 C.E.

1. Background

In 1565, Pieter Bruegel the Elder painted the frigid northern European landscape in his work *Hunters in the Snow*, one of a series of winter landscape paintings (Kemp, 2008). That year was just one of many years during the 16th century when winters in Europe were particularly severe. Historical and physical records from many parts of the world indicate cooler temperatures for much of the period between about 1450 and 1850 A.D (PAGES 2k consortium, 2013). The proposed reasons for this period of cooler temperatures, the Little Ice Age (LIA), have varied with region and include solar variability, particularly the Maunder Minimum of sunspots (Eddy, 1976; for review see Lean, 2010), periods of strong and frequent tropical volcanic eruptions (Miller et al., 2012; Schurer et al., 2014), declining Northern Hemisphere summer insolation associated with the long cycles of the Earth’s orbital parameters (Kauffman et al., 2009), and land use/land cover changes (He et al., 2014). The LIA was preceded by a period of warmer temperatures, the Medieval Climate Anomaly (MCA) from about 950 to 1250, although this period exhibited much more heterogeneity in the timing and regionality of the responses (Bradley et al., 2003; Diaz et al., 2011).

The last millennium has a rich archive of annually-dated records that give us a longer perspective on climate variability and change than the instrumental period (see Jones et al., 2009 and PAGES 2k consortium, 2013 for reviews). An extensive network
of tree ring records provide measures of the year-to-year as well as longer-term variability of continental temperatures at mid- to high latitudes (e.g., Fritts et al., 1979; Briffa, 2000; Cook et al., 2006) and continental moisture at low latitudes (e.g. Cook et al., 1999; 2010) over the last millennium, although the spatial coverage decreases at earlier times. Similarly, ice cores in Greenland and Antarctic (e.g., Vintner et al., 2009, Graf et al., 2002) and Arctic lake records (e.g., Kauffman et al., 2009) contribute to the reconstruction of high-latitude regional temperature variability over the last millennium. At lower latitudes, analyses of stalagmites of caves establish monsoon variability (e.g., Wang et al., 2005) and coral records have been employed to assess centennial to millennial variability of the El Nino-Southern Oscillation (e.g., Cobb et al., 2003; 2013).

Because of its high data density, the last millennium is an excellent time period over which to quantify the relative importance of natural and anthropogenic forcings in explaining recent and more distant changes to climate. For this reason, it has caught the attention of policy makers and has been featured prominently in the Intergovernmental Panel on Climate Change (IPCC) reports. For the Coupled Model Intercomparison Project Phase 5 (CMIP5) and as a contribution to the IPCC Fifth Assessment (AR5), modeling groups worldwide completed simulations for the period 850-1850 (hereafter referred to as the Last Millennium [LM]) with the same models and using the same resolutions as for future projections (Taylor et al., 2012). The forcing protocols were defined by the Palaeoclimate Modelling Intercomparison Project (PMIP) and included several scenarios for solar and volcanic forcings to allow testing of the structural uncertainties in different reconstructions (Schmidt et al., 2011). Eleven modeling groups completed the CMIP5 LM simulations, but the available LM
simulations remain somewhat limited in utility. The GISS and CSIRO modeling groups considered the sensitivities to different combinations of forcings (Phipps et al., 2013; Schmidt et al., 2014). Only the CSIRO model completed an ensemble of simulations (3) to assess the role of internal variability on the regional signals imprinted in the proxy data.

The CESM Last Millennium Ensemble (CESM-LME) provides a resource for the research community for understanding and attributing the climate variations of the last 1156 years to internal variability and both natural and anthropogenic forcings. The scientific questions that motivated our project include the pressing need to evaluate the ability of models such as CESM to capture observed variability on multidecadal and longer time-scales, to determine the characteristics (i.e. intensity, spatial structure, seasonality) of variability associated with the individual natural forcings versus purely internal variability, and to permit a longer-term perspective for detection and attribution studies.

2. Experimental design

The CESM-LME employs version 1.1 of CESM with the Community Atmosphere Model version 5 [CESM1(CAM5); Hurrell et al., 2013], the same model as the CESM Large Ensemble (CESM-LENS, Kay et al., 2015) except for resolution of the atmosphere and land components. The CESM-LME uses ~2-degree resolution in the atmosphere and land components and ~1-degree resolution in the ocean and sea ice components (1.9x2.5_gx1v6). At this resolution, we can simulate ~25 years per day on the NWSC (NCAR-Wyoming Supercomputing Center) Yellowstone computer, and complete a full simulation from 850 to 2005 in ~45-50 days.
Before starting our CESM-LME, we spun up the model for an 1850 control simulation of 650 years, from which an 850 control simulation was branched and run for an additional 1356 years (Fig. 1). All LME simulations were started from year 850 of the 850 control simulation (i.e. after 200 years of the 850 control). The only difference among ensemble members was the application of small random round-off (order $10^{-14}$ °C) differences in the air temperature field at the start of each ensemble member. Both control simulations were run in excess of 1000 years to overlap with the LME simulations, allowing removal in our analyses of any trends still present. In surface temperature and top-of-atmosphere (TOA) net incoming flux, these trends are on the order of 0.02°C per century and 0.01 W m$^{-2}$, respectively. For sea ice area, there is no significant trend in the control simulation. Over the period from 850 to 1849, the global mean ocean temperature in the 850 control integration cools by only ~.004°C per century, reflecting a continuing adjustment toward equilibrium in the deepest ocean levels below ~500m. The global mean ocean salinity freshens by ~5 x $10^{-5}$ PSU per century, indicating an insignificant adjustment between global freshwater reservoirs.

The choices of LM forcings and their implementations follow those used in our LM simulation with CCSM4 (see Landrum et al., 2013 Fig. 1 and its discussion for more details). The forcings over this period include orbital, solar, volcanic, changes in land use/land cover and greenhouse gas levels. We adopt the concentrations of the well-mixed greenhouse gases (CO$_2$, CH$_4$, and N$_2$O) from high-resolution Antarctic ice cores (Schmidt et al., 2011) and calculate of the seasonal and latitudinal distribution of the orbital modulation of insolation from the equations in Berger (1978). For the volcanic forcing, we adopt the ice core-based index of Gao et al. (2008) and version 1 of their
modeled estimates of loadings as a function of latitude, altitude, and month (see http://climate.envsci.rutgers.edu/LVI2/ for corrections that have been made since to this dataset). Stratospheric aerosols are prescribed in the CESM as a fixed single-size distribution in the three layers in the lower stratosphere above the tropopause. Changes in total solar irradiance (TSI) are prescribed using the Vieira et al. (2011) reconstruction on which an estimated 11-yr solar cycle has been imposed and spectral solar irradiance derived using linear regression of TSI at each spectral interval (see Schmidt et al., 2011 for details). We merged the Pongratz et al. (2008) reconstruction of land use with that of Hurtt et al. (2011). The only plant functional types (PFTs) that are changed are those for crops and pasture; all other PFTs remain at their 1850 control prescription.

For the continuation of the LME simulations from 1850 to 2005, we adopted the same forcings as the LENS with the exception of including orbital changes in insolation not considered in the LENS. The LME-adopted ozone and aerosol forcings are fixed at the 1850 control values until 1850 and then include the evolving anthropogenic changes to 2005.

The LME forcings are detailed in Fig. 2. Volcanic events of varying strength have been a recurring feature of the last millennium (Fig. 2a) with the largest eruptions in terms of total global stratospheric volcanic sulfate aerosol injection occurring in 1258 (Samalas, 257.9 Tg), 1452 (Kuwae, 137.5 Tg), and 1815 (Tambora, 109.7 Tg). These can be compared to the 1991 Pinatubo eruption with a total global stratospheric volcanic sulfate aerosol injection estimated to be 30.1 Tg (Gao et al., 2008). Periods of frequent, large tropical eruptions occurred in the last half of the 13th century and the first half of the 19th century. It should be noted that there are major differences in both
the timing and magnitude of volcanic forcing between the two suggested PMIP3 reconstructions (Gao et al., 2008; Crowley and Unterman, 2013) as shown in Schmidt et al. (2011) arising from different methodologies and uncertainties inherent in reconstructing volcanic deposition events from polar ice cores. Solar variability is characterized by a pronounced quasi-11-year sunspot cycle in insolation of about 1 W m$^{-2}$, a variability that itself changes with time. Sunspot counts back to 1610 and longer, indirect records of solar activity from ice cores and tree rings indicate periods of ~70-100 years with much reduced solar activity. These include the Wolf (~1280-1350), Sporer (~1460-1550), and Maunder (~1645-1715) grand solar minima, with an increase in the TSI of about 0.1% from the Maunder Minimum to today (Fig. 2b). Concentrations of major greenhouse gases have been largely stable prior to the late 19th century, with only relatively small variations related to natural feedbacks in the carbon and nitrogen cycles. Major increases in CO$_2$, N$_2$O, CH$_4$ and F$_{11/12}$ occur during the 20th century (Fig. 2c). Over the past two centuries, there has also been a substantial increase in crop and pasture extent, amounting now to nearly a third of all land regions (Fig. 2d). When the magnitudes of these forcings are estimated from the net TOA tropical clear-sky flux over ocean (Fig. 2e), the decadal-mean variability associated with volcanic events (several W m$^{-2}$) dominates that of other sources (orbital and LULC contributions not shown for clarity and due to their small net forcing). Solar variability acts as a secondary source of variability (~0.2 W m$^{-2}$), an influence that for example can be discerned from 1650-1750 (Fig. 2e).

We have completed 30 CESM1(CAM5) simulations for the LME project. These include an ensemble of 10 simulations with all forcings as well as smaller ensembles.
with each forcing separately (Table 1, Fig. 1). To assess the influence of anthropogenic-forced changes in ozone and aerosols, we completed two simulations starting at year 1850 of the control simulation and continuing to 2005. One full-forcing ensemble member (#7) saved 6-hourly atmospheric output for forcing regional models and extremes analyses, and one full-forcing ensemble member (#10) included the simulation of radiocarbon in the ocean for comparison to related proxies. We also extended one full-forcing ensemble member (#2) to 2100 with RCP8.5 forcing and completed the CMIP6 abrupt 4xCO$_2$ and 1% to 4xCO$_2$ experiments.

### 3. Results

In this section, we illustrate some examples of the capabilities and applications of the LME. Importantly, we compare to proxy reconstructions, where possible, to provide an important first benchmark for validating CESM1(CAM5).

#### a. Surface temperature

Northern Hemisphere (NH) mean surface temperature has been estimated from numerous proxy reconstructions (Mann et al. 2009, Ammann and Wahl 2007, Moberg et al. 2005, Juckes et al. 2007) and is compared to the ensemble mean of LME full-forcing runs (Fig. 3). The simulated NH surface temperature confirms the reconstructed “hockey stick”–like pattern with generally warm conditions near 1100 (coincident with the Medieval maximum of TSI and only weak volcanic activity) and a gradual cooling until the 19th century. The LME resolves cooling events associated with major volcanic eruptions (1258, 1452, 1815), and sustained intervals of cool temperatures during the late 13th century, from the early 17th to mid 18th centuries, and during the early 19th
The most notable feature of both the simulated and reconstructed records is the large warming evident since the late 19th century, the timing and magnitude of which differs across reconstructions but whose overall magnitude is well-observed in the instrumental period and agrees well with the LME simulations.

The LME full-forcing simulations are able to capture the overall globally warmer conditions during the MCA (950-1250) relative to the LIA (1450-1850) present in the proxy record; global average temperature differences range from 0.12 to 0.17°C (Fig. 4). As shown in Fig. 2, the most significant differences in radiative forcing between the MCA and LIA are the very large volcanic eruptions during the LIA as compared to only weak volcanic activity during the MCA. The resultant response shows polar amplification of the simulated temperature responses, particularly in the Arctic (~3-4X the global temperature change), more muted temperature changes in the tropics, and greater temperature changes over the continents than the oceans. The LME full-forcing simulations compare favorably overall, both qualitatively and quantitatively, with the annual temperature reconstruction of Mann et al. (2009). The polar regions, where the simulations indicate the largest temperature anomalies between the MCA and LIA, are not reported in the reconstruction due to the challenges involved in validating proxy records.

Regionally, the strong reconstructed warming at MCA as compared to LIA over northern Europe is simulated in some, but not all, ensemble members and illustrates the importance of an ensemble modeling approach. None of the ensemble members reproduce the cooler SSTs of the eastern tropical Pacific or the much warmer SSTs in
the tropical Atlantic during the MCA as compared to LIA present in the Mann
reconstruction.

Over North America, all the LME full-forcing simulations exhibit warmer
conditions during the MCA than the LIA though underestimate the magnitude of surface
temperature differences as compared to the Mann reconstruction (Fig. 4 and 5). The
much higher surface temperatures during the MCA as compared to the LIA are only
robust regionally across the full-forcing ensemble for Hudson Bay and the Canadian
Arctic Archipelago. Each of the individual LM forcings contribute to this pattern of a
warmer MCA than LIA in the Canadian Arctic, but only the volcanic-only and orbital-only
simulations consistently so (though not always above the 95% confidence criteria). For
solar-only forcing and GHG-only forcing, not all ensemble members reproduce the
pattern of warming in the Canadian Arctic. These results illustrate that for attribution of a
climate response to individual forcings during the LM, ensembles of simulations are a
requirement.

In our simulations, the late 20th century warming cannot be solely explained by
increased solar irradiance during the 20th century (Scafetta and West, 2008). Fig. 6
shows annual surface temperature anomalies, for present-day (PD; 1950-2000) minus
LIA (1450-1850), for the single-forcing and full-forcing LME runs. For each forcing
scenario, ensemble members with minimum and maximum global temperature
differences are shown. The increased GHG at the end of the 20th century result in
strong simulated warming over the entire globe except over the North Atlantic region.
The simulated small cooling in the latter region is associated with a weakening of the
simulated AMOC in the 20th century. Cooling due to strongly increased Northern
Hemisphere aerosols during the 20th century counteracts approximately half of the simulated GHG warming in the Northern Hemisphere. The 20th century aerosol loadings in the Southern Hemisphere are smaller and as such the simulated cooling associated with the aerosols is much less than in the Northern Hemisphere. Sea ice plays an important feedback in the model responses to the GHG and aerosol increases during the 20th century (see next section). Increases in solar irradiance from the Maunder Minimum to the late 20th century (~0.1% in our reconstruction) force a more modest, consistent warming over the Arctic Ocean and North America. The other forcing - volcanic, orbital, and land use - have much less effect on the surface temperature changes simulated for the last half of the 20th century as compared to the LIA.

b. Sea ice

Sea ice has been declining rapidly over recent decades (e.g., Stroeve et al., 2008), with large contributions from both forced anthropogenic trends and internal variability (Swart et al., 2015). Prior to the second part of the 20th century, sea ice records are sparse, but the existing reconstructions suggest that the recent sea ice loss has been unprecedented during the last 1400 years (e.g., Kinnard et. al, 2011). By looking at the different LME single forcing simulations compared to the 10-member full forcing LME, we can begin to attribute features in the full forcing simulation to individual forcings over the last millennium, allowing us to put the recent changes into a longer-term context.

In agreement with the sea ice reconstructions for the last millennium by Kinnard et al. (2011), the largest signal in the sea ice extent in the LME simulations is the strong
decline in the NH sea ice extent over recent decades (Fig. 7). By looking at the single-forcing simulations, it is clear that this decline is driven by GHG. In the NH, the sea ice decline in the GHG-only simulations already starts in 1850, but other forcings, in particular the ozone-aerosol forcing and the volcanic forcing, delay and reduce the magnitude of the realized sea ice extent decline in the fully-forced simulations. In the SH, the effect of the GHG forcing on sea ice is offset less than in the NH, and the simulated sea ice extent decline in the SH shows the GHG forcing impact very strongly. Prior to the decline due to GHG, sea ice extent had been slowly increasing (Fig. 7), in agreement with the slow cooling over that period (Figure 3) and with sea ice reconstructions (e.g., Massé et al., 2008). Superimposed on the slowly increasing sea ice prior to 1850, some strong volcanic forcing events are clearly visible in the fully-forced ensemble mean, e.g. for the NH around 1825 and 1260. Furthermore, it is clear that significant interannual variability has occurred over the last millennium, making it important to employ ensembles to investigate the detailed climatic impact of sea ice changes.

c. Precipitation and drought

Anthropogenic forcing is expected to lead to large changes in 21st century hydroclimate, with most models projecting drying trends over North America (Wuebbles et al. 2014). Additionally, proxy records indicate that decadal to multidecadal 'megadroughts' are quite common in many locations (Woodhouse & Overpeck 1998); using the LME, we can gain insight into the drivers of such events and the degree to which anthropogenic effects may dominate.
Toward that end, Fig. 8 shows the difference in Palmer Drought Severity Index (PDSI) between 1950-2000 and 850-1850, and PDSI differences from the North American Drought Atlas tree ring reconstruction (Cook et al. 2004). The LME full-forcing ensemble (Fig. 8a) shows overall drier conditions in the western United States and Northern Mexico during the 20th century, with wetting in the Midwest and Eastern US. Single forcing runs suggest this is due primarily to greenhouse gas and ozone/aerosol effects (Fig. 8c,d), with minor contributions from other forcings. Agreement with the NADA is quite good over Mexico and the eastern US, but some discrepancies exist in the West (Fig. 8e).

d. Climate variability

Modes of internal of climate variability affect regional patterns of temperature and precipitation. The influence of forcings on both global mean and lateral gradients in radiative fluxes provides a means by which they may interact with these modes. While various reconstructions have suggested multidecadal to centennial modulation of these modes by past forcings (e.g. Trouet et al., 2009; Li et al., 2011; Knudsen et al., 2011), the extent to which natural forcings over the past millennium have acted as a pacemaker for these modes remains an open but important question for constraining near-term future projections and improving risk estimates.

The spectra of two key modes of variability are shown in Fig. 9, where the LME simulations are differentiated into groups that do not incorporate volcanic forcing (red), that include only volcanic forcing (green) and that incorporate all forcings (black) along with their associated 1-sigma range (shaded). A role for volcanic forcing in increasing the power of the Atlantic Multidecadal Oscillation (AMO, defined as the area-weighted
North Atlantic SST anomalies from 0° to 60°N and 80°W to 0°E after removing the global [60°S to 60°N] SST anomalies) at low frequency is suggested as the ensemble mean power of both the all-forcing and volcanic forcing runs lies consistently above the range of power in the distribution of runs without volcanic forcing. Differences in the power exhibited by the Niño3.4 index (defined as the area-weighted monthly SST from 5°S to 5°N and 120° to 170°W after removing the long term monthly means) are also considerable at various bands, however, the spread amongst ensemble members is also considerable and at no frequency is the distribution of either the all forcing or volcanic forced runs outside of the spread of spectra evident in the control simulation and LME runs without volcanic forcing.

Previous model simulations indicate a positive correlation of the AMO with the Atlantic Meridional Overturning Circulation (AMOC) although the lead/lag relationship varies with model and forcing (Delworth and Mann, 2000; Ottera et al., 2010). The CESM-LME AMOC, NH annual mean sea ice extent (SIE) and AMO all show significant responses to the seven largest volcanoes for 30/12/20 years after an event, respectively (Fig. 10). Composite response shows that the AMO cools immediately, whereas sea ice increases markedly for 2 years and then decreases continuously for another ~10 years. AMOC strength increases – and the increase of the AMO two years post-event is consistent with significant correlations with the AMOC (Danabasoglu et al., 2012). The AMOC strengthens for ~14 years after an event, then gradually decreases to pre-event levels 30 years post-event. We also found that while both NH SIE and the AMO respond similarly to medium sized volcanic eruptions (~4 to -10 W m² reduction in solar flux), the AMOC does not.
The El Niño/Southern Oscillation (ENSO) accounts for the majority of interannual climate variations, and is expected to respond to external forcing through a variety of feedback processes (Collins et al. 2010). The available proxy data suggests a possible increase in ENSO variance over the last millennium (e.g. Cobb et al. 2013, McGregor et al. 2010), although the magnitude of the increase is fairly modest. Modulations of ENSO by external forcings in the LME are investigated in Fig. 11. Consistent with the lack of response in the overall Niño3.4 power spectrum, little systematic change is seen during epochs of varying solar irradiance and the running variance does not respond to volcanic eruptions. Some ensemble members do show a trend toward higher Niño3.4 variance during the 20th century, but even this behavior is not consistent among all realizations of the 850-2005 period. Panels b and c show a stacked coral δ¹⁸O record from the tropical Pacific (Cobb et al. 2013) and the LME Niño3.4 running variance restricted to only periods during which coral data exists, respectively; the trends estimated from such gappy records are highly variable, suggesting that not only internal variability but data availability strongly limits our ability to estimate anthropogenic influences on ENSO strength.

The Mount Tambora eruption of April 1815, one of the largest eruptions during the historical period, caused both local devastation and widespread human and climate impacts for several years following. The subsequent year became known as “the year without a summer” because of the unusually cold and wet summer conditions in North America and Europe (Stommel and Stommel, 1979, 1983; Stothers, 1984) that led to poor harvests and famine. It is the third largest eruption in terms of stratospheric aerosol forcing since 850 yielding an anomalous reduction in TOA clear-sky solar flux
over the tropical ocean of more than \(-4 \text{ W m}^{-2}\) in the decadal mean (Fig. 2e) and a peak monthly reduction in excess of \(-27 \text{ W m}^{-2}\).

Past studies using historical observations and proxy data have argued that large tropical volcanic eruptions lead to an El Nino-like warming in the post eruption period (Handler, 1984; Adams et al., 2003; McGregor et al. 2010; Wahl et al., 2014), but these results remain controversial (Self et al., 1997; Robock, 2000). Fig. 12 illustrates the benefit of using an ensemble approach to study the relationship between ENSO and large tropical eruptions. Individual realizations of the cold season tropical Pacific sea surface temperature anomalies one year after the Tambora eruption peaks are shown in Fig. 12 as well as the ensemble mean of the 15 ensemble members with volcanic forcing. The ensemble mean, as well as 9 of the individual realizations exhibit an El Nino-like warming in the Eastern Pacific. This is double the average likelihood for an El Nino to occur in any given cold season. The other 6 ensemble members simulate cooler or no change in the cold season tropical surface temperature anomalies one year after the Tambora eruption.

4. Summary, next steps, and community involvement

The CESM-LME provides a more comprehensive look at climate responses since 850 C.E. than previously available to the community. Our initial analyses of the CESM-LME highlight the importance of an ensemble approach to investigate the detailed climate responses chronicled by the proxies. That said, the present ensemble does not completely account for uncertainties in the magnitudes of forcing factors; we chose one of the possible reconstructions for each of the forcings of the LM (Schmidt et al., 2011) for our simulations. Alternate reconstructions have since become available. For land
use, the reconstruction of Kaplan et al. (2011) estimates total global land use-land cover change at 1850 A.D. to be approximately twice as large as that in the Hurtt et al. (2011) or Pongratz et al. (2008) reconstructions. New volcanic reconstructions incorporating additional records and better dating suggest that volcanic aerosol loading for some of the largest eruptions (e.g. Samalas (1257) and Kuwae (1453) may have been overestimated by 20-50%, and others underestimated by 20-50% (Sigl et al., 2014). In addition, the magnitude of solar variability is still debated (Schmidt et al., 2012; Schurer et al., 2014). As well, our simulations do not include the ‘top-down’ effect of solar variability (Meehl et al., 2009). We plan to complete additional simulations to more fully explore the role of the uncertainties in the reconstructed forcings, as well as simulations with the high-top chemistry version of CESM, the Whole Atmosphere Community Climate Model (WACCM), to also explore the climatic responses to the stratospheric ozone changes to the solar intensity variations.

This paper provides a few examples of the responses of CESM1(CAM5) to the natural and anthropogenic forcings from 850 to 2005. For further analyses by the community, the CESM-LME outputs are publicly available via the Earth System Grid (https://www.earthsystemgrid.org) as single variable time series in self-documenting lossless compressed netCDF-4 format. High-frequency output for regional modeling and analysis of extremes is available for ensemble member #7 of the full forcing simulations. The CESM-LME web page (https://www2.cesm.ucar.edu/models/experiments/LME) provides more background on the CESM-LME project, including diagnostic plots, lists of publications and ongoing projects, and instructions for reproducing the simulations.

Acknowledgments:
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References:


Wahl, E. R., H. F. Diaz, J. E. Smerdon, C. M. Ammann, 2014: Late winter temperature response to large tropical volcanic eruptions in temperate western North


**Table 1. CESM-LME simulations.** Additional information about the simulations including the forcing datasets, saved variables, diagnostics, model support and known issues can be found at the CESM-LME webpage:

http://www2.cesm.ucar.edu/models/experiments/LME

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Fig. 1. Details of the initial states and simulation lengths of the CESM-LME control and forced runs.

Fig. 2. Evolution of the major forcings used for the CESM-LME including a) volcanic mass, b) total solar irradiance (TSI), c) greenhouse gas concentrations (GHG), d) crop and pasture extent, and e) TOA net clear sky shortwave flux over the tropical oceans.

Fig. 3. Northern Hemisphere annual surface temperature anomalies from full forcing runs (black) with 1-σ range (shaded) versus various reconstructions and instrumental observed (HADCRUT2V). A thirty-year Gaussian smoothing has been applied. Reconstructions plotted are: B2000 (Briffa, 2000), BOS2001 (Briffa et al., 2001), ECS2002 (Esper et al., 2002), HCA2006 (Hegerl et al., 2006), JBB1998 (Jones et al., 1998), MBH1999 (Mann et al., 1998), MJ2003 (Mann and Jones, 2003), MSH2005 (Moberg et al., 2005), O2005 (Oerlemans, 2005), PS2004 (Pollack and Smerdon, 2004), and RMO2005 (Rutherford et al., 2005).

Fig. 4. Annual surface temperature anomalies, MCA (950-1250) minus LIA (1450-1850), from Mann et al. (2009) proxy-based reconstruction (top) and as simulated in ten CESM-LME full-forcings simulations. Stippling indicates differences not statistically significant at the 95% level using the student-t test.
Fig. 5. Annual surface temperature anomalies, MCA minus LIA, over North America, as simulated in the CESM-LME full and single-forcing simulations. Stippling indicates differences not statistically significant at the 95% level using the student-t test. See Fig. 4 for definition of time periods.

Fig. 6. Annual surface temperature anomalies, PD (1950-2000) minus LIA (1450-1850), as simulated in CESM-LME. Shown for each forcing set are the ensemble members with minimum and maximum global average temperature differences. Stippling indicates differences not statistically significant at the 95% level using the student-t test.

Fig. 7. Annual mean sea ice extent for the Northern (a) and Southern (b) Hemisphere in the CESM-LME single-forcing and the full-forced ensemble simulations (smoothed with a 20-year running mean). The ensemble spread in the full-forced ensemble is shown by grey shading in all panels and the ensemble mean of the 10 fully-forced ensemble members is shown in cyan in the full forcing panels.

Fig. 8. Palmer Drought Severity Index composite differences between 1950-2000 and 850-1850 in the LME (panels a, c-h) and observations (North American Drought Atlas; Cook et al. (2004), panel b). Model PDSI is computed using the Penman-Monteith potential evapotranspiration method; negative values indicate drier conditions for 1950-2000 relative to 850-1850, positive values indicate wetter conditions. Stippled locations show no significant difference between the 1950-2000 and 850-1850 periods, using a two-tailed T test.
Fig. 9. Ensemble-mean AMO and Niño3.4 power spectra and associated standard deviation range for full-forcing runs (black), all runs without volcanic forcing (red) and runs with only volcanic aerosols (green). The time period used is 850-1849, and thus excludes the ozone-aerosol forcing runs. The 850 control run is included in the “without volcanic forcing” set.

Fig. 10. Lead-lag relationships of AMOC PC1, NH annual sea ice extent (SIE), and AMO composite responses to seven volcanic events with largest impacts on annual radiation in the North Atlantic (greater than -10 W m$^{-2}$ reduction in solar flux) for CESM-LME all-forcing and volcanic-only simulations. Composites for individual simulations/ensemble of simulations are shown in light grey/black. The response is shown as a deviation from the 30-year mean prior to the event. Year 0 is the first year of reduction in solar flux. Dashed lines represent 2 standard deviations for individual (light grey) and ensemble (black) bootstrap random events. The AMOC is represented with the leading order principal component (PC) time series and EOF analysis for 33°S-60°N.

Fig. 11. Analyses of modulations in ENSO variability over the Last Millennium. a) 20-year running Niño3.4 variance computed from the full-forcing CESM-LME simulations; green and yellow shaded regions indicate periods of minima and maxima in solar insolation, respectively. Horizontal dashed lines show the 10th and 90th percentiles of Niño3.4 variance from the 850 control simulation. b) Same variance data as a), plotted only over time periods for which coral proxy data have been collected. c)
Variance of coral oxygen isotope records ($\delta^{18}O$) from a stacked time series, constructed using data from Palmyra (Cobb et al. 2003), Maiana (Urban et al. 2000), Nauru (Guilderson & Schrag 1999), Tarawa (Cole et al. 1993), and Christmas (McGregor et al. 2011).

Fig. 12. Tambora eruption (April 1815) and the simulated tropical Pacific surface temperature anomalies during winter 1816. The December to February (DJF) seasonal surface temperature anomalies for each simulation with volcanic forcing shown here are computed relative to each simulation’s long term annual cycle.
Fig. 1. Details of the initial states and simulation lengths of the CESM-LME control and forced runs.
Fig. 2. Evolution of the major forcings used for the CESM-LME including a) volcanic mass, b) total solar irradiance (TSI), c) greenhouse gas concentrations (GHG), d) crop and pasture extent, and e) TOA net clear sky shortwave flux over the tropical oceans.
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