

Past climate in the Baltic Sea region

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The climate in the Baltic Sea region during the last millennium

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Abstract

Variability and long-term climate change in the Baltic Sea region is investigated for the pre-industrial period of the last millennium. For the first time dynamical downscaling covering the complete millennium is conducted with a regional climate model in this area. As a result of changing external forcing conditions the model simulation shows warm conditions in the first centuries followed by a gradual cooling until c. 1700 before temperature increases in the last centuries. This long-term evolution, with a Medieval Climate Anomaly (MCA) and a Little Ice Age (LIA), is in broad agreement with proxy-based reconstructions. However, the timing of warm and cold events is not captured at all times. We show that the regional response to the global climate anomalies is to a strong degree modified by the large-scale circulation in the model. In particular, we find that a positive NAO-phase simulated during MCA contributes to enhancing winter temperatures and precipitation in the region while a negative NAO-anomaly in the LIA reduces them. In a second step, the regional ocean model RCO is used to investigate the impact of atmospheric changes onto the Baltic Sea for two 100 yr time slices representing the MCA and the LIA. Besides the warming of the Baltic Sea the water becomes fresher at all levels during the MCA. This is induced by increased runoff and stronger westerly winds. Moreover, the oxygen concentrations in the deep layers are slightly reduced during the MCA. Additional sensitivity studies are conducted to investigate the impact of even higher temperatures and increased nutrient loads. The presented experiments suggest that changing nutrient loads may be more important determining oxygen depletion than changes in temperature or dynamic feedbacks.

1 Introduction

The climate of the last millennium is characterised by large long term variability. Warmer conditions prevailed in large parts of the Northern Hemisphere at the beginning of the last millennium (950–1250 AD) with colder conditions thereafter

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(1400–1700 AD) (e.g. Mann et al., 2009; Ljungqvist et al., 2012). These periods are known as the Medieval Climate Anomaly (MCA) and the Little Ice Age (LIA) whereas the exact period varies between different studies. Differences between periods prior to 1850 reflect internal unforced variability and changes in external forcing as man-made contributions were small before that. Most studies reveal that such long term variability can be triggered by the sun (e.g. Haigh, 1996; Guiot et al., 2010) since the sun’s irradiation varies over time scales from days to millennia (Lean, 2005). Another source of influence comes from strong volcanic eruptions. These can considerably reduce hemispheric mean temperature over several years (Timmreck et al., 2009). Moreover, model studies indicate that natural unforced long-term variability could have significantly contributed to form the exceptional periods of the MCA and the LIA (Mann et al., 2009; Jungclaus et al., 2010).

Regional temperature variability in Europe is related to changes in the North Atlantic Oscillation (NAO) (Hurrell, 1995). Proxy data (Mann et al., 2009; Trouet et al., 2009) and model studies (Gómez-Navarro et al., 2011) indicate that a positive NAO phase prevailed during the MCA whereas the NAO was negative during the LIA (Spanghehl et al., 2010). However, the influence of the NAO on the Baltic Sea region is limited (Kauker and Meier, 2003). Nevertheless, for NAO anomalies it is known that the location and intensity of storm tracks is affected (e.g. Spanghehl et al., 2010), leading to anomalies in precipitation. Here, results based on tree rings show diverging results in Northern Europe for the key periods of the MCA and the LIA. For instance, Jönsson and Nilsson (2009) reveal evidence that large parts of the LIA (namely 1560–1590 and 1657–1675 AD) were drier whereas Helama et al. (2009) points out that it was wetter from 1220 until 1650 AD.

Model studies focusing on the climate evolution of the last millennium have so far generally been conducted with global climate models (GCMs) even if they sometimes focus on certain regions (Gouirand et al., 2007). This limits the comparison of model results with proxies as tree rings, which are affected by regional/local features that might not be represented appropriately in GCMs. Dynamical downscaling is one possibility

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to increase the resolution and include local effects into the model results. However, the finer resolution of regional climate models (RCMs) demands more computer power and restricts its application. As a consequence, no downscaling approach exists for Fennoscandia covering the entire millennium so far. A long integration of in total 600 yr has been done with an earlier version of RCA3 although only at relatively coarse resolution (Graham et al., 2009). For the Iberian Peninsula a first dynamical downscaling approach was conducted for the whole millennium by Gómez-Navarro et al. (2011). They found good agreement between their modelled temperature (used model is RCM MM5) and proxy-based reconstructions. We present in this study first results of a dynamical downscaling approach including most of Europe, whereas we focus on the Baltic Sea region.

Moreover, the downscaled atmospheric conditions are used to simulate the response of the Baltic Sea to changing forcing conditions with the Rossby Centre Ocean model (RCO). The Baltic Sea is a large estuary sensitive to changes in the environmental conditions. For instance, it is known from sediment studies that the deep bottom layers were hypoxic during the MCA in contrast to the LIA (Zillén and Conley, 2010). The period of hypoxia during the MCA correlate to population growth and large-scale changes in land use, and Zillén and Conley (2010) argue that human perturbation was an important factor contributing to hypoxia during the MCA similarly as in the 20th century (Conley et al., 2009; Hansson and Gustafsson, 2011; Gustafsson et al., 2012). Following their argumentation salinity and temperature anomalies cannot explain hypoxia during the MCA. Therefore, it resulted from anthropogenic impacts (Zillén et al., 2008; Zillén and Conley, 2010). However, historical nutrient loads are not estimated yet and the effect of pure natural climate change during the MCA has not been investigated so far.

Hansson and Gustafsson (2011) used the PROBE-Baltic ocean model to simulate Baltic Sea conditions during the last 500 yr. Their results indicate a strong decrease of bottom oxygen only after 1950 with no significant signals before including the LIA. They suggest that increased nutrient inputs most likely dominate the expansion of hypoxic

bottom areas during the last century. This is in agreement with other shorter reconstruction starting in 1850 (Gustafsson et al., 2012) and 1902 (Meier and Kauker, 2003), respectively. However, little is known about the pristine state of the Baltic Sea earlier during the last millennium.

In this study, we investigate climate variability in the Baltic Sea and its surrounding area for the pre-industrial period of the last millennium with the MCA and LIA as key periods. We present the first regional climate simulation for this area covering the full millennium. Variability of atmospheric parameters connected to the forcing and internal variability are discussed. Here, we focus on parameters relevant for the state of the Baltic Sea as temperature, precipitation, and wind speed. The large-scale atmospheric circulation is compared to that derived from proxy-based reconstructions of the NAO-index. Results of the ocean model are used to highlight to what degree these changes effect the Baltic Sea.

The paper is organised as follows: in Sect. 2 the models and the experimental setup is described. Section 3 focus on changes of the atmosphere whereas in Sect. 4 effects on the Baltic Sea are investigated. The results are summarised and discussed in Sect. 5.

2 Models and experiments

2.1 Model description

2.1.1 The Rossby Centre regional climate model

The Rossby Centre Regional Climate model version 3 (RCA3, Samuelsson et al., 2011) is based on the numerical weather prediction model HIRLAM (Unden et al., 2002). In the present setup it operates on a rotated longitude-latitude grid with a horizontal resolution of 0.44° (approx. 50 km). Here, RCA3 is run with 24 vertical levels and a time step of 30 min. The model domain with its 102×111 grid boxes is shown in Fig. 1. RCA3

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has a tiled surface scheme including a detailed snow storage parametrisation over land. Lake surfaces are simulated with a lake model whereas the sea surface temperatures (SSTs) are taken from the driving GCM.

RCA3 has been used and verified for many applications ranging from paleo (Kjellström et al., 2010; Strandberg et al., 2011) to future simulations (Kjellström et al., 2011; Nikulin et al., 2011). The model is in good agreement with observations if it is forced with ECMWF re-analysis-40 (ERA40, Uppala et al., 2005) boundary conditions. The small biases include a too high MSLP in the Mediterranean region, a warm bias in the northeastern part, a cold bias in Southern Europe and North Africa, and an underestimation of the diurnal temperature cycle (Samuelsson et al., 2011). The finer resolution contributes to a better captured geographical distribution of precipitation in RCA3 than in ERA40 (Lind and Kjellström, 2009; Samuelsson et al., 2011). For a more detailed model description and validation we refer to Samuelsson et al. (2011).

In this study, the driving global climate model is ECHO-G (Legutke and Voss, 1999). ECHO-G consists of the spectral atmospheric GCM ECHAM4 and the coupled ocean-thermodynamic sea ice model HOPE-G. ECHO-G has been used and evaluated for many paleo climate studies (e.g. Zorita et al., 2005; Kaspar et al., 2007). The internal generated variability of the NAO is similar to the variability in observations, and the correlation of the NAO with precipitation and the 2m air temperature is adequate (Min et al., 2005).

2.1.2 The Rossby Centre ocean model

We use the Rossby Centre Ocean model (RCO) with a domain that covers the entire Baltic Sea (Fig. 1). RCO is a Bryan-Cox-Semtner primitive equation circulation model with a free surface (Killworth et al., 1991) and open boundary conditions (Stevens, 1991) in the Northern Kattegat. It has a horizontal resolution of two nautical miles (about 3.7 km). The 83 vertical layers are equally thick with a layer thickness of 3 m. The baroclinic and barotropic time steps amount to 150 and 15 s, respectively. The sea surface heights (SSHs) at the open boundary in the Northern Kattegat between

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Denmark and Sweden is based on SLP gradients over the North Sea taken from the RCA3 simulation. In the case of inflowing water temperature and salinity are nudged towards climatological profiles. For a detailed model description the reader is referred to Meier et al. (2003) and Meier (2007).

RCO is forced with 10 m wind, 2 m air temperature, 2 m specific humidity, precipitation, total cloudiness and sea level pressure fields from RCA3. The SSH and river runoff are calculated based on the RCA3 data following Meier et al. (2012).

2.2 Experimental setup

RCA3 is forced with an ECHO-G simulation starting 7000 BP (Wagner et al., 2007; Hünicke et al., 2010). The ECHO-G model is forced with variations in orbital parameters, solar irradiance and greenhouse gases between 7000 BP and 1998 AD. Since the focus in this study is on the last millennium we use boundary conditions from 950 AD until 1998 AD to force RCA3. A more elaborative description of the ECHO-G Holocene simulation can be found in Hünicke et al. (2010).

Figure 2 shows the evolution of the CO₂ concentration and the solar insolation as prescribed for ECHO-G and RCA3. The variability of the CO₂ concentration is relatively small in the pre-industrial period. Until 1850 the CO₂ concentration varies between 270 and 290 ppmv. A rapid increase occurs thereafter which is known to be of anthropogenic origin. Towards the end of the simulation the CO₂ concentration reaches values of more than 360 ppmv.

Variations in solar radiation appear during the whole millennium. The most distinct minima were the Oort, the Spörer, the Maunder, and the Dalton minimum whereas each minima lasts over several decades (Fig. 2b). The solar variability is scaled to an insolation difference between present day and the Maunder Minimum of 0.3%, as estimated by Lean et al. (1995).

In a next step, the downscaled atmospheric data are used to force RCO. Four experiments are performed in this study (Table 1). For the MCA as well as for the LIA 100 yr periods are selected with exceptional high and low temperatures according to the RCA

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simulation (see Sect. 3, Fig. 3). We force RCO with output from RCA for which a bias correction had to be applied for temperature and wind speed. Biases arise partly from the large-scale atmospheric circulation in the driving GCM (cf. Kjellström et al., 2011). Further, some biases over the Baltic Sea arise from the prescribed SSTs taken from ECHO-G. Due to the coarse scale of the ocean in the global model is the representation of conditions in the Baltic Sea area rather low. The bias correction was made as a change in the mean value by adding spatial variable fields for each month. The correction fields are derived from a comparison with an RCA simulation forced with ERA40 for the period 1961–2008. Note that the variability within the RCA3 simulation remains unchanged by adding constant fields.

In addition to the simulations forced directly with the RCA3-data (referred to as RCO-MCA and RCO-LIA), two sensitivity experiments for the MCA were performed. An increase of 2 K for the 2 m air temperature was implemented constant in time and space to amplify the temperature effect whereas all other parameters stayed unchanged. Since the total temperature difference compared to the LIA is about 3 K in this case we refer to this simulation as RCO-3K. In the other experiment higher nitrate and phosphor loads were used (RCO-HL). While all other simulations use nitrate and phosphate concentrations in the river runoff as estimated by Savchuk et al. (2008) for 1850 AD we increase these values by roughly 30 % for RCO-HL.

We used estimations of the 1850 state of the Baltic Sea as starting conditions. These are taken from other simulations and establish the best estimation available for the pre-industrial time. The same starting field is used for RCO-MCA and RCO-LIA. The sensitivity studies are initialised from the RCO-MCA run, more precisely the final state in 1299, to reduce spin-up effects. However, since a spin-up must be assumed for all simulations we skip the first 50 yr and use only the second half of the simulations.

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3 Results: atmospheric response

3.1 Temperature evolution

Figure 3 illustrates the temporal evolution of the mean temperature in the Baltic Sea area during the pre-industrial period (950–1900 AD) as 50 yr running means. In addition to the annual mean (black), winter (DJF, blue) and summer (JJA, red) temperatures are illustrated as anomalies w.r.t. the pre-industrial mean. The warmest period appears in the 13th century whereas the 17th century is the coldest in the simulation (Fig. 3). These centuries are referred to as MCA and LIA in this study and are highlighted in Fig. 3.

The absolute maximum in the Baltic Sea region is reached towards the end of 13th century with a winter temperature anomaly of 0.85 K, whereas the coldest winters are about 0.81 K colder in the 17th century than on average (Fig. 3). These values reflect the largest amplitudes within the MCA and LIA centuries whereas the mean signals covering the entire century are weaker (+0.41 and -0.47, respectively, Table 2). Moreover, the signals are most pronounced during the winter season whereas the summer months indicate a much weaker difference between the MCA and LIA.

The anomalies in the annual mean are mainly driven by changes in winter temperatures since these have the larger amplitude. For example, the cold peaks in the 17th century are connected to winter anomalies only while there is no clear signal in any other season (Fig. 3). Therefore, the amplitude of annual means is higher than the amplitude for summer means and lower than the amplitude of winter means (Table 2). For 50 yr periods the maximum temperature difference between the MCA and the LIA is 0.87 K.

The evolution of the temperature anomalies follows for all seasons to some degree the fluctuations in the solar and GHG forcing (cf. Fig. 2). Especially, the high solar irradiation in the 12th and 13th century as well as low irradiation in the Maunder minimum agree well with the temperature evolution. However, as expected, there are differences and all details are not correlated as the local and regional temperature is not identical

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to the large-scale global and hemispheric evolution. Moreover, the model-generated internal variability is not necessarily in phase with that in the real climate system. Discrepancies include that the lowest insolation appears in the Spörer minimum whereas the coldest temperature occur in the Maunder Minimum where on the other hand the GHG concentrations are somewhat lower. Moreover, the evolution of the temperature is characterised by more peaks than the forcing series.

An additional minimum is found around 1810 AD. Here, once more the strongest signal can be found especially for the winter running mean. This minimum seems to be connected with the solar Dalton minimum (Fig. 2b) even if it occurs a little later. Most climate models simulate only a clear Dalton minimum if volcano eruptions are considered (Wagner and Zorita, 2005). However, in the forcing of both ECHO-G and RCA3 volcanoes are not included. Therefore, it cannot be excluded that the signal is to a great extent connected to internal variability in our simulation. On the other hand, it is also possible that small forcings amplify through positive internal feedbacks (e.g. ice albedo feedback) on the regional scale.

The long-term evolution of the temperature climate is in broad agreement with proxy-based reconstructions of the climate (e.g. Ljungqvist et al., 2012) with a warm Medieval climate anomaly and a cold Little Ice Age. However, details between reconstructions and the model integration differ. An example is the fact that the model simulated maximum is in the 12th and 13th centuries while the reconstructions have this maximum earlier. As the internal variability in the model simulation is not in phase with that in the real climate system we do not perform any direct comparisons with proxy data in detail. Instead, we look at the variability in the model and a proxy-based reconstruction of wintertime temperatures from Stockholm (Leijonhufvud et al., 2010, Fig. 4b). The comparison reveals that the variability in RCA3 is slightly higher on both inter-annual time scales (standard deviation 2.1 K in RCA3, 1.8 K in the proxy-record) and longer time scales (0.64 K/0.43 K on 30-yr filtered data). Further, in the next section we compare the relation between variables in the model to the same relation in proxy-based reconstructions.

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3.2 The North Atlantic oscillation

3.2.1 NAO-temperature relationship

The state of the NAO controls the climate of Northern Europe to a large degree (Hurrell, 1995). Especially, the winter temperature in the Baltic Sea area and precipitation in the Norwegian mountains is affected. The positive correlation of the NAO-index and winter temperatures in Stockholm in the model is shown in Fig. 4a. Positive NAO phases correspond to positive temperature anomalies in Stockholm whereas winters with low NAO index have below normal temperatures on average. The highest correlation can be found for 20-yr running means as shown in Fig. 4a. A similar picture is revealed when proxy-based reconstructions of temperature (Leijonhufvud et al., 2010) and NAO-index (Luterbacher et al., 2002) are compared to each other (Fig. 4b). A detailed analysis of the correlation between the two series is performed for the pre-industrial time period 1543–1900. The period is chosen so that a number of years with missing data in the temperature proxy is not included. Before calculating correlation coefficients both temperature and NAO series are linearly detrended to remove the contribution from any long-term trend to the correlation. The correlation coefficient for RCA3 is 0.52 calculated on an inter-annual basis while the corresponding number is 0.47 for the two proxy series. Applying a 20-yr running mean to the data increases the correlation coefficients to 0.66 in RCA3 and 0.52 in the proxies (Fig. 4). These results show that the model has a somewhat stronger dependence on the NAO for the winter time temperatures in Stockholm than indicated by the proxies. This holds true both on inter-annual and decadal time scales but on even longer time scales the opposite is the case as the corresponding correlation coefficient reduces to 0.59 in RCA3 while being higher, 0.68, in the proxies when a running 30-yr mean filter is applied.

Although the results show a relatively strong correlation between the NAO index and the temperature climate in the Baltic Sea region, the NAO index only explains some 25–50 % of the total variability on different time scales. For the remainder other factors needs to be taken into account. However, it is evident that internal variability in the

model, partly manifested as variations in the NAO index, has a strong influence on the regional and local climate. Thereby, we note that it cannot be expected that anomalies in the imposed external forcing will have a strong impact on the climate in a small area like the Baltic Sea region.

3.2.2 Evolution of the NAO index

Most proxy and model studies agree that the LIA was characterised by prevailing negative NAO conditions (e.g. Luterbacher et al., 2002; Spangehl et al., 2010). For the MCA the confidence level is not that high since very few proxy data sets reach that far back in time. In general, proxy based results depend strongly on the applied method which includes a certain amount of uncertainty (e.g. Bürger and Cubasch, 2005) so that results should be compared advisedly. Nevertheless, some reconstructions exist that could be used. Trouet et al. (2009) find evidence that the phase with the highest NAO during the last millennium appeared during the MCA (Fig. 5). According to this study more than two centuries (12th and 13th) were characterised by an exceptional high winter NAO index. Even if the mean state of the NAO is positive during the MCA in our simulation a comparable strong and robust signal as in that particular proxy-based record is not reproduced. Two periods with very high simulated NAO indices occur within the MCA, namely towards the end of the 12th century and around 1300 AD (Fig. 4). Nevertheless, these two maxima are separated by a deep minimum with a negative NAO index situation, a feature that is absent in the Trouet et al. (2009) proxy data until the 15th century. Here, it should be noted that these large scale circulation features in the RCM are steered mainly by the driving GCM. The RCA model domain is too small to differ considerably from ECHO-G in the NAO phase. That ECHO-G is not reproducing the positive NAO phase is in-line with other model studies of the MCA (Mann et al., 2009; Jungclaus et al., 2010). Reasons for that are widespread and reach from a missing signal in the Pacific where a strong La Nina like pattern was prevailing influencing Europe via teleconnections (Mann et al., 2009) to the lack of a fully resolved stratosphere which can modify the solar signal response down to the surface (e.g. Gray et al., 2010).

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Moreover, as discussed above, an absolute match cannot be expected since the development of the NAO comprises to a large degree internal variability which is supposed to be independent from the external forcing. Therefore, an overall agreement of proxy data and model results cannot be requested.

After 1500 AD more proxy series of the NAO-index exist. The data of Luterbacher et al. (2002) and Cook et al. (2002) are included in Fig. 5 to demonstrate the uncertainties related to proxy data. Gómez-Navarro et al. (2011) discussed already for the Luterbacher et al. (2002) and Trouet et al. (2009) data that there is no agreement in general. Partly, similar signatures can be found as for instance the minimum before 1800 which is also shown by the Cook et al. (2002) data. However, large parts are less correlated. While the Luterbacher et al. (2002) data show a distinct minimum during the LIA without any strong positive anomalies, the other series include several maxima with positive anomalies. Also, the strong positive NAO anomaly during the MCA in the Trouet et al. (2009) data indicates that colder than average conditions should have been prevailing in Greenland and parts of Northern Canada which is not the case following temperature reconstructions for that area (Ljungqvist et al., 2012). Obviously, there are large uncertainties included in the proxy data and its interpretation. Therefore, it remains an open question if the NAO was that high and constant during the MCA as indicated by the Trouet et al. (2009) data.

3.3 Runoff and precipitation

The runoff is estimated from the net water budget (precipitation minus evaporation) over the Baltic Sea drainage area using a statistical model (see Meier et al., 2012 for more information). It is illustrated in Fig. 6 in combination with precipitation averaged over the catchment area. The 50-yr running means of the two variables show considerable multi-decadal variability in a close relationship. The correlation coefficient of both 50 yr running means is 0.68, whereas the correlation coefficient for annual values is 0.66.

Runoff and precipitation are considerably larger during the MCA than the LIA. The 50-yr peak runoff during the MCA is approximately $800 \text{ m}^3 \text{ s}^{-1}$ (+6%) higher than in

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the LIA. Moreover, the mean values for the runoff over the full 100-yr periods (+2.4%) and the last 50 yr (+3.8%) differ also considerably between the MCA and the LIA. The results here, with more precipitation during the MCA compared to the LIA, is in general agreement with the findings in Graham et al. (2009) although a detailed comparison is difficult to make as they are only focusing on the northerly drainage basins of the Baltic Sea area. Moreover, the fluctuations in precipitation are in agreement with modelled changes in the NAO since the largest positive precipitation anomalies occur during winter whereas there is no significant difference during summer months (not shown).

3.4 Spatial anomaly pattern

As mentioned above, most proxy and model studies indicate that warm and cold anomalies as the MCA and LIA are connected with high or low NAO index situations – at least during winter. We derive from our simulation that the variability of the NAO contributes especially for the extremes on the multi decadal time scale whereas other parameters have an effect on even longer time scales (Sect. 3.1). Consistently, the SLP difference between the full MCA and LIA shows a rather weak negative NAO pattern which is mainly characterised by lower pressure in the North (Fig. 7). The pressure gradient between Iceland and Portugal is lower by more than 1 hPa.

The pattern becomes much stronger for the 50 yr periods. A clear positive NAO anomaly is simulated for the MCA in comparison to the LIA. In contrast to the usual NAO signature which shows the strongest SLP reduction over Iceland (e.g. Hurrell, 1995) the simulated signal is shifted towards Northern Scandinavia. Nevertheless, the north-south pressure gradient is higher all over Europe and by more than 2.75 hPa between Iceland and Portugal. A positive NAO predominate also during autumn though with a reduced amplitude. The differences are negligible in summer and spring (not shown).

Since we defined our MCA and LIA periods by warm and cold phases in the running mean (Fig. 3) we see a clear signal in the spatial pattern of the temperature differences. In correspondence to stronger winter amplitudes in the running mean we find

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the largest differences in the spatial winter pattern (Fig. 7). The temperature difference increases from 0.5 K at the southern tip of the Baltic Sea to more than 1.5 K in Northern Fennoscandia for the full 100-yr periods. While smaller differences exist for spring and autumn there is almost none in summer (not shown).

For the 50 yr periods the winter mean temperatures over Scandinavia differ between 1 and 2.5 K over Fennoscandia (Fig. 7). The amplitude increases from the Southwest towards the Northeast. For the shorter period an increase can be seen also during summer (roughly 0.4 K, not shown).

Since the differences in the SLP patterns are only minor for the full periods large signals cannot be expected for the 10 m wind speed. A slight increase is found for the MCA with local maxima of 0.2 ms^{-1} over the Baltic Sea (Fig. 7). Compared to the mean wind speed of $5\text{--}7 \text{ ms}^{-1}$ that change is not much.

The signal is stronger over the Southern Baltic Sea for the 50 yr periods in line with the higher NAO index. The mean wind speed is enhanced by more than 0.2 ms^{-1} in large areas and by up to 0.4 ms^{-1} over the Southern Baltic proper. A similar pattern is prevailing for the gustiness but with a higher amplitude (more than 0.5 ms^{-1} , not shown). These differences are in agreement with the SLP signatures. Moreover, changes in the SLP field assume that the westerly component is mainly strengthened during the MCA.

4 Results: Baltic Sea response

4.1 Evolution of water temperature, sea ice and salinity

The atmospheric data of the 13th and 17th century are then used to simulate the behaviour of the Baltic Sea in the MCA and the LIA, respectively. The evolution of temperature, salinity and maximum ice cover are shown in Fig. 8 for both extreme periods for the last 50 yr.

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As a consequence of the atmospheric boundary conditions the volume averaged temperature of the Baltic Sea is higher during the MCA than during the LIA. The mean difference adds up to 0.50 K which is less than the averaged air temperature signal over the Baltic Sea (0.73 K). The remaining energy is consumed at least partly for the melting of sea ice. The annual variability is that high that some years out of the MCA are colder than warm years of the LIA.

The mean maximum ice cover is somewhat smaller during the MCA than in a run forced with ERA40 at the boundary (Table 3). This is reasonable since the MCA is supposed to be approximately as warm as the recent warm period (Ljungqvist et al., 2012). The ice cover is higher during the LIA. The mean difference between the MCA and the LIA is about 44 000 km².

In comparison with observations collected by the Finnish Meteorological Institute (Seinä and Palosuo, 1993) two differences are evident. First, the present model setup seems to underestimate the maximum annual ice cover by more than 20 000 km². Second, the long-term variability (50-yr average) within the observational data is larger than the difference between the modelled MCA and LIA conditions. This adds further evidence that the simulated temperature difference between the MCA and LIA is underestimated.

The mean difference in salinity between RCO-MCA and RCO-LIA is 0.69 PSU (Fig. 8c). This is an effect of both the increased runoff (Fig. 6) and the stronger westerlies (Fig. 7). First, the increased runoff dilutes the Baltic Sea surface water. Then, due to internal recycling saltwater inflows are diluted as well which reduces the amount of salt entering the Baltic Sea (Meier, 2005). In addition, the increase of the mean westerly wind speed causes anomalous high sea level in the Baltic Sea. This produces a barotropic pressure gradient and hampers the flow of high-saline water through the Danish Straits into the Baltic proper (Meier and Kauker, 2003). A similar long term sensitivity for stronger westerly winds for the Baltic Sea was shown by Zorita and Laine (2000).

4.2 Vertical profiles in the Gotland Deep

In this section, vertical profiles at the Gotland Deep are investigated (Fig. 1). This location lies in the middle of the Baltic Sea and is representative for large parts of it. For instance, due to the strong stratification the bottom layers are poorly ventilated, and only strong major inflows of salty water from the North Sea renew this bottom water intermittently.

Temperature differences between RCO-MCA and RCO-LIA are quite uniform over the entire water column and amount to 0.5 K in the mean. The differences are larger in the mixed layer down to 50 m (0.68 K) and somewhat smaller around 60 m (0.29 K), whereas the deep layers warm by about 0.5 K (Fig. 9a).

Salinity reductions amount to 0.59 PSU at the surface and 0.78 PSU in the layers below the halocline. The largest decrease appears in the halocline with a reduction of more than 1 PSU. This implies in combination with the temperature response that the mixed layer is deeper in the MCA than in the LIA. This is in line with results of future projections, where the Baltic Sea becomes warmer and fresher as well (Meier et al., 2011a).

The oxygen profiles show the typical behaviour with high and saturated concentrations close to the surface and lower concentrations in deeper layers. Compared to observations the concentrations are higher in the deep layers and neither anoxic nor hypoxic conditions occur in the MCA or the LIA. Nevertheless, some differences can be identified between the two periods. First, a small reduction of oxygen concentrations (-0.1 ml l^{-1}) can be seen in the mixed layer. This is determined by higher temperatures and reflects the reduction of the saturation level. Second, a slight increase ($+0.3 \text{ ml l}^{-1}$) is present around 60 m depth which reflects the deepening of the mixed layer which brings higher oxygen concentrations into deeper waters. Finally, oxygen concentrations are slightly reduced in the bottom water in the MCA compared to the LIA. Maximum decrease is found close to the bottom with a reduction of 0.5 ml l^{-1} .

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4.3 Sensitivity studies

Proxy studies suggest that temperature differences between the MCA and LIA have been stronger than simulated in our model setup (e.g. Leijonhufvud et al., 2010). Hence, a sensitivity experiment is performed with 2 K higher air temperatures compared to RCO-MCA.

The first order effect is an increase of 1.3 K in the volume averaged temperature compared to RCO-MCA (Fig. 8a). This difference is constant over the entire simulation, whereby the annual variability is the same in RCO-MCA and RCO-3K. The difference between the change in the driving temperature and the response of the water is explained by energy consumption due to melting sea ice and enhanced evaporation triggered by higher temperatures. The difference between MCA-3K and RCO-LIA adds up to 1.84 K for the volume averaged temperature of the Baltic Sea. The vertical temperature profile at Gotland Deep reveals a similar behaviour for RCO-MCA and RCO-3K only with higher amplitudes for RCO-3K. For instance, the difference in SST in comparison with RCO-LIA is enhanced to more than 2 K. Moreover, a minimum is still existing around 60 m depth.

The mean maximum sea ice cover is strongly reduced and amounts to less than 100 000 km² in RCO-3K. However, an ice cover exists during all winters with a minimum extension of 24 500 km² (Fig. 8b). Salinities are somewhat higher (+0.1 PSU) in RCO-3K than in RCO-MCA, whereby the annual variability is very similar in both simulations (Fig. 8c). The main reason for this is likely a higher evaporation due to higher temperatures. However, a long lasting effect of the different starting points cannot be excluded completely. The absolute salinity difference to RCO-LIA is still a reduction but less pronounced (−0.58 PSU). The shape of anomalies in the vertical profile is comparable to the RCO-MCA–RCO-LIA differences.

The oxygen profile for RCO-3K indicates a clear reduction for the mixed layer oxygen concentrations. The reduction by roughly 0.4 ml l^{−1} reflects the effect of reduced saturation levels due to higher temperatures. The signal is clearly emphasised in comparison

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to RCO-MCA. However, the difference to RCO-MCA becomes smaller with increasing depths and there is hardly any difference left in the bottom layers.

In addition to studies reporting on higher temperature differences between the MCA and the LIA other studies report on higher nutrient loads during the MCA mainly caused by the heyday of the surrounding population and their agriculture (Zillén et al., 2008). This motivates another sensitivity experiments with higher loads (RCO-HL) while the residual forcing is the same as in RCO-3K.

Since nutrient loads do not have distinct feedbacks on the physics the response in e.g. temperature and salinity is not different from MCA-3K. Therefore, the developments (Fig. 8) and profiles (Fig. 9a,b) are identical in both sensitivity experiments for these variables. The major influence of increased nutrient loads is on the oxygen concentrations in the deep layers (Fig. 9c). Whereas the oxygen concentrations in the mixed layer are the same for both sensitivity experiments it is obvious that higher loads increase the oxygen consumption in deeper layers. The mean oxygen concentration in the bottom layer of the Gotland Deep is 3.3 ml^{-1} in comparison to 4.5 ml^{-1} in RCO-LIA and 4.0 ml^{-1} in RCO-MCA and RCO-3K, respectively. Therefore, mean oxygen concentrations are considerably reduced in RCO-HL but the model does not simulate hypoxic or anoxic conditions.

5 Discussion and conclusions

For the first time a dynamical downscaling approach has been performed for the complete last millennium over Northern Europe. This is so far a unique approach to high-light regional features in the context of natural climate variability. Moreover, the high resolution of the downscaled atmospheric data opens the possibility to perform the first simulations with a dynamical regional ocean model for the Baltic Sea covering the MCA. We demonstrate that the atmospheric model setup of ECHO-G and RCA3 produces a relative warm period in the early centuries of the millennium and cold conditions in the 17th century which is in general agreement with reconstructions (Ljungqvist

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et al., 2012) and other model studies (e.g. Mann et al., 2009). While most of the long-term change is driven by the external forcing, changes on shorter periods (up to few decades) are connected to internal variability, including anomalies in the NAO. Neither the external forcing nor the signal in the NAO can explain the temporal evolution of the temperature independently. Herewith, both amplitude and timing of the modeled MCA and LIA are considerably affected by multi-decadal variability. Pronounced temperature anomalies on century-long time scales are not simulated. This is in agreement with the results of other model studies (Mann et al., 2009; Jungclaus et al., 2010). The temperature differences between the simulated warmest and coldest 50-yr periods are rather small especially for the summer if compared to proxy indicated signals (Ljungqvist et al., 2012) while the variability in winter is as high, or even higher, than that retrieved from a proxy-based reconstruction of Stockholm temperatures. The low variability in summer could be a result of the missing volcanic forcing. There is indication from model studies that the inclusion of volcanic eruptions strengthens the agreement with proxy data (Wagner and Zorita, 2005; Jungclaus et al., 2010). Further, the use of SSTs from the GCM simulation for the Baltic Sea can dampen the temperature signals considerable.

Moreover, also in broad agreement to proxy-based reconstructions our simulation shows that a positive NAO-phase was prevailing during the MCA (Mann et al., 2009), whereas the LIA was characterized by a negative NAO phase (Spanghehl et al., 2010). However, at least the signal for the MCA is weaker and not as consistent as reconstructed (Trouet et al., 2009). As elucidated in the result section, it remains unclear to what extent this is a missing signal in the model setup or the results from the proxy data are overrated. Nevertheless, we demonstrate that the difference in the NAO phase contributes to the positive winter temperature anomaly during the MCA as well as higher westerly wind speed and more precipitation. As a result, the Baltic Sea becomes fresher in RCO-MCA similar to future scenario studies (Meier et al., 2011b). Oxygen concentrations are only slightly effected and show a reduction in the mixed layer and below the halocline, whereas they increase in the range of the halocline owing to the enlarged mixing depth. The sensitivity experiments point out that the reduction in the

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deep layers is not initiated by higher temperatures, but are sensitive to the amount of nutrient loads. As only the nutrient load concentrations are kept constant, the higher runoff leads in total to a raise of the loads in RCO-MCA. According to this, we find a stronger reduction of oxygen concentrations in the sensitivity experiment RCO-HL. Here, the oxygen concentrations are reduced by 1.3 ml l^{-1} compared to RCO-LIA. In addition to effects of nutrient loads, changes in stratification have an impact on oxygen concentrations in the deep. For instance, oxygen conditions improve in the deep due to a better ventilation initiated by reduced salinity (Hansson and Gustafsson, 2011) or higher wind speed (Meier et al., 2012). These factors counteract the oxygen reduction in RCO-MCA and the sensitivity studies. Another source for oxygen rich water for the deep layers is inflowing water from the North Sea (Matthäus and Franck, 1992). However, there is no indication in our simulations that the frequency of major inflows differs between RCO-MCA and RCO-LIA. Other effects as discussed for scenario simulations, for instance increased decomposition/oxidation rates of organic matter due to higher temperature (Meier et al., 2011a), are not relevant because the pools are empty according to the low loads.

Anyway, the reduction of bottom oxygen concentrations does not lead to a spread of hypoxia or anoxia during the MCA in the model as it is found in sediment studies (Zillén and Conley, 2010). In fact, the oxygen conditions are simply too high to develop any hypoxia. This holds true also for the sensitivity study with 30 % higher nutrient loads which is perhaps a rather high estimation taking the size of the medieval population into account.

Therefore, further investigations are needed to explain the reconstructed increase of hypoxia during the MCA. Basically, three shortcomings of this study have been identified. Improved atmospheric forcing conditions including the feedbacks from volcanic activity could be crucial. Moreover, estimations of the nutrient supply during the MCA would be helpful (Zillén and Conley, 2010). Finally, potential shortcomings in the parametrization of oxygen consumption in the bottom layers under nutrient-poor conditions have to be improved as discussed also by Gustafsson et al. (2012).

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Table 1. Overview of performed RCO simulations for the MCA and LIA. The overview includes the simulated periods, the forcing 2 m air temperature anomaly relative to the LIA state (T_0), and the nutrient loads.

	Period	Temperature	Nutrient loads
RCO-LIA	1600–1699	T_0	1850 level
RCO-MCA	1200–1299	$T_0 + 1\text{K}$	1850 level
RCO-3K	1200–1299	$T_0 + 3\text{K}$	1850 level
RCO-HL	1200–1299	$T_0 + 3\text{K}$	1850 level +30%

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Table 2. Overview over modelled temperature anomalies during the MCA and the LIA. Anomalies are shown for winter (DJF), summer (JJA), and the annual mean (ann). The rows show values for the entire century, the second half of the century, and the absolute maximum (minimum) for a 50 yr period within the MCA (LIA).

	MCA DJF	LIA DJF	MCA JJA	LIA JJA	MCA ann	LIA ann
Entire century	0.41	−0.47	0.10	−0.25	0.22	−0.37
2nd half of the century	0.60	−0.78	0.07	−0.24	0.23	−0.47
Abs. max./min. for 50 yr	0.85	−0.81	0.19	−0.30	0.40	−0.47

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Table 3. Mean annual maximum sea ice extent (10^3 km^2) modelled for RCO-LIA, RCO-MCA, RCO-3K and a simulation forced with ERA40 boundary conditions (1961–2000). In addition, observed maximum sea ice cover for the period 1720–2000 is presented (Seinä and Palosuo, 1993). Here, we show the absolute minimum (1902–1951) and maximum (1766–1815) over 50 succeeding years as well as the mean over the period 1961–2000.

	RCO-MCA	RCO-LIA	RCO-3K	ERA40
model results	149	193	96	162
	1902–1951	1766–1815		1961–2000
Observations	182	264		185

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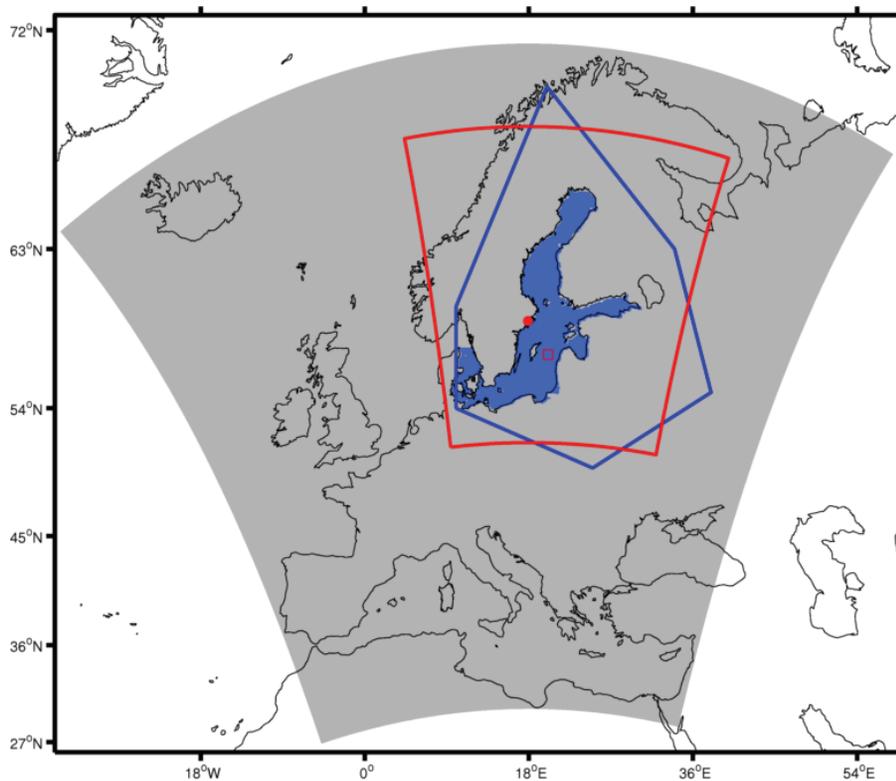


Fig. 1. The model domains of RCA (grey) and RCO (blue) are highlighted together with specific locations. The location of the temperature reconstruction side by Leijonhufvud et al. (2010) is marked with a red point whereas the location of the Gotland Deep is indicated by a square. The framed areas indicate the averaging areas for the Baltic region temperature (red) and the catchment area (blue).

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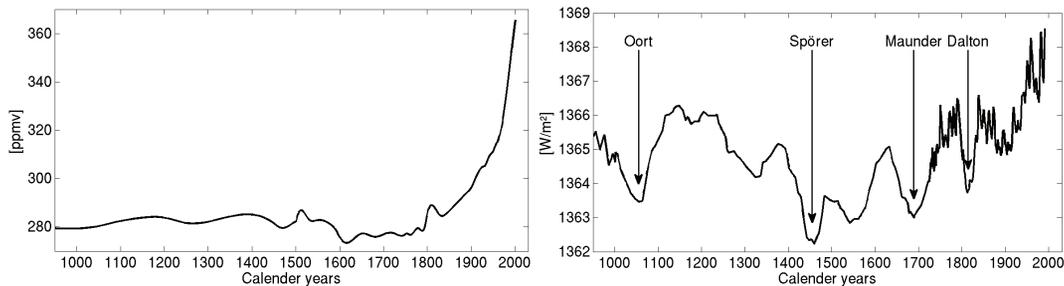


Fig. 2. The main forcing parameters of the simulation. Prescribed CO₂ concentrations (ppmv) (left panel) and solar irradiance (W m⁻²) (right panel).

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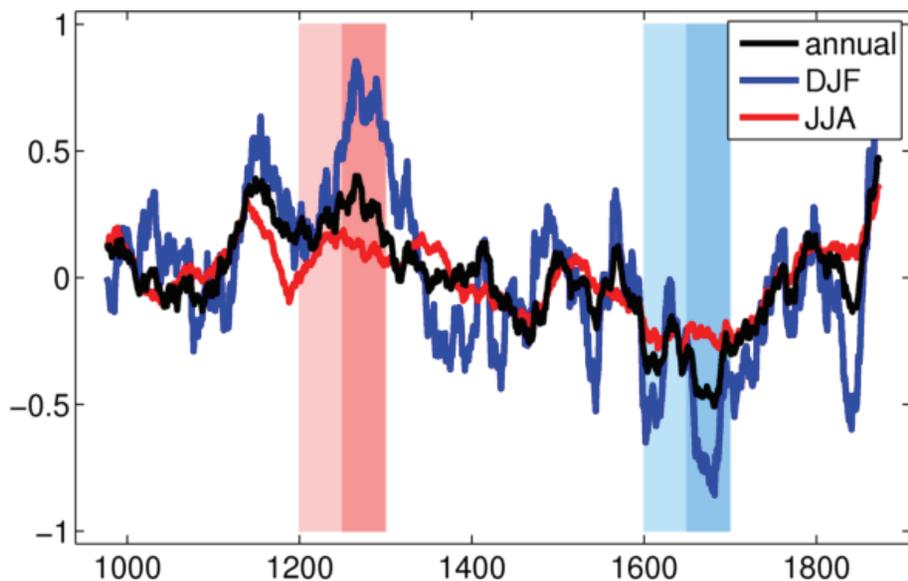


Fig. 3. The 2 m-temperature anomaly w.r.t. the pre-industrial mean (950–1900) is illustrated for the winter (DJF), summer (JJA) and annual mean averaged over the Baltic Sea region. The coloured sections highlight the periods that are defined as MCA (red) and LIA (blue) in this study. The darker colours reflect the 50 yr periods which are considered for the analysis of the Baltic Sea.

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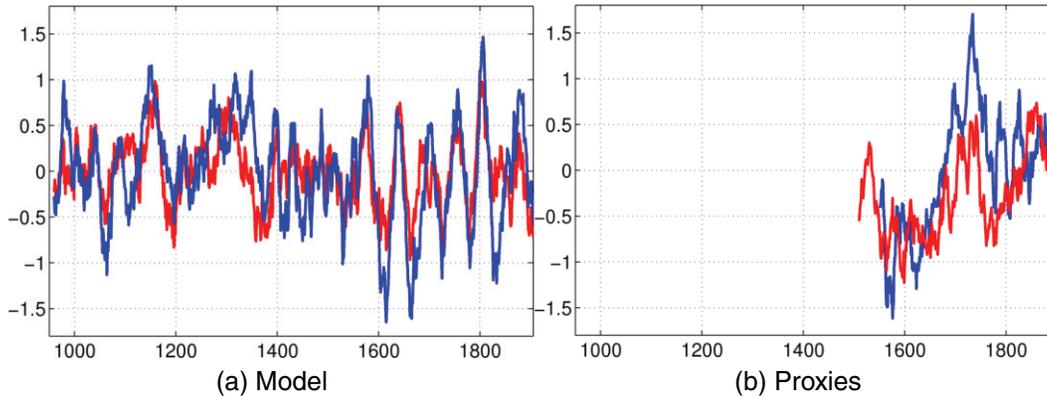


Fig. 4. (a) The simulated NAO index (red) and the 2m-temperature for a grid box representing Stockholm (blue). (b) Proxy-based reconstructions of the NAO-index (Luterbacher et al., 2002) and 2m-temperature in Stockholm (Leijonhufvud et al., 2010). The modelled NAO index is computed according to Hurrell (1995) for DJF whereas the temperature reflects DJFM values for both modelled and reconstructed series. All series are shown as 20-yr running means.

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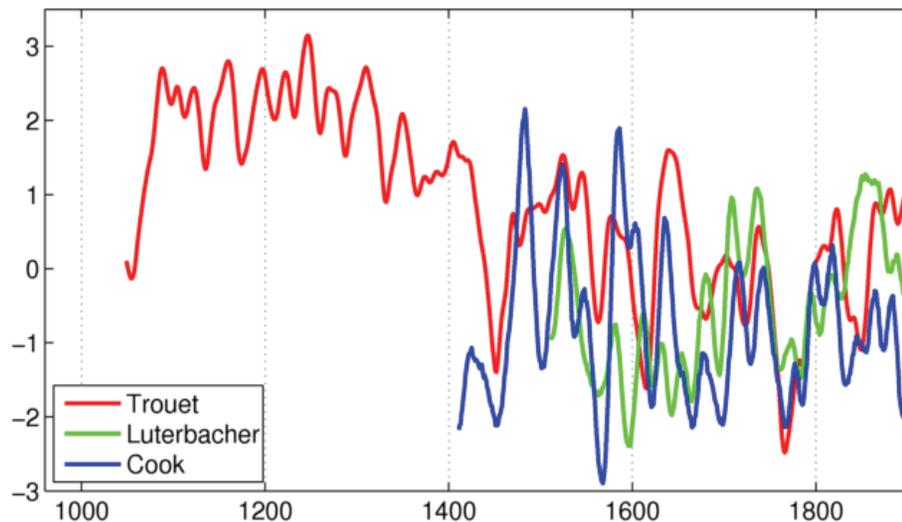


Fig. 5. The winter NAO index for proxy data (Cook et al., 2002; Luterbacher et al., 2002; Trouet et al., 2009). All time series are low-pass filtered with a cut-off frequency of 30 yr and normalised thereafter.

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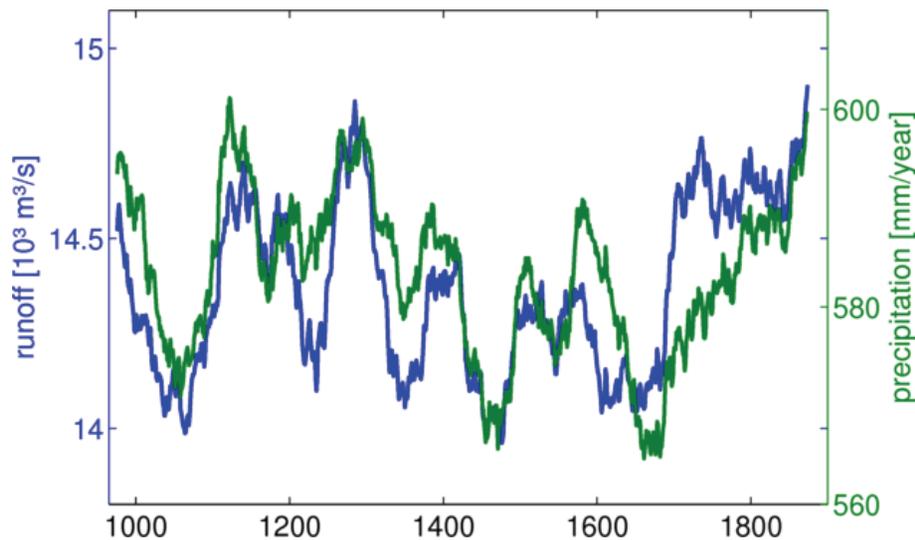


Fig. 6. Precipitation in the Baltic Sea catchment area (green, see Fig. 1) and statistical estimated runoff (blue) as 50-yr running means.

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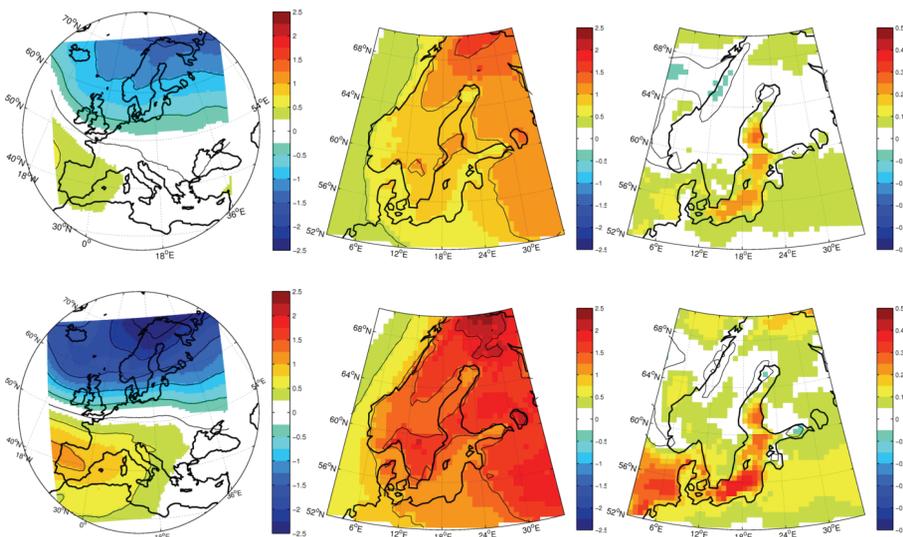


Fig. 7. SLP (left panel), temperature (middle panel) and wind speed (right panel) differences between the 100 yr (top panel) and the 50 yr (bottom panel) periods of the MCA and LIA as indicated in Fig. 3a). All variables show winter (DJF) means.

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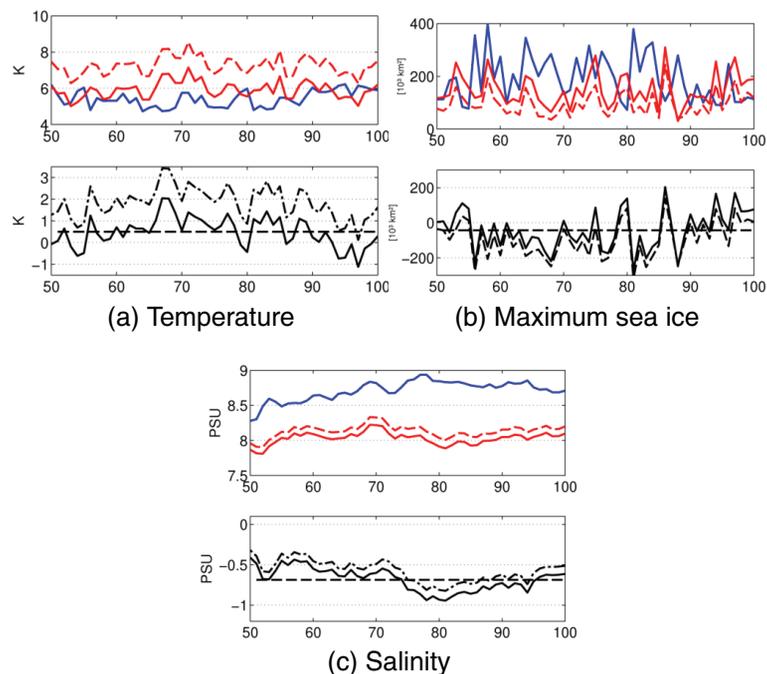


Fig. 8. Volume averaged annual mean temperature **(a)**, maximum sea ice extent **(b)**, and volume averaged annual mean salinity **(c)**: RCO-LIA (blue solid lines), RCO-MCA (red solid lines), RCO-3K (red dashed lines), RCO-MCA minus RCO-LIA (black solid lines), and RCO-3K minus RCO-LIA (black dash-dotted lines). The straight lines indicate mean differences. The x-axis indicates simulated years of the 13th and 17th century for RCO-MCA, RCO-3K, and RCO-LIA, respectively.

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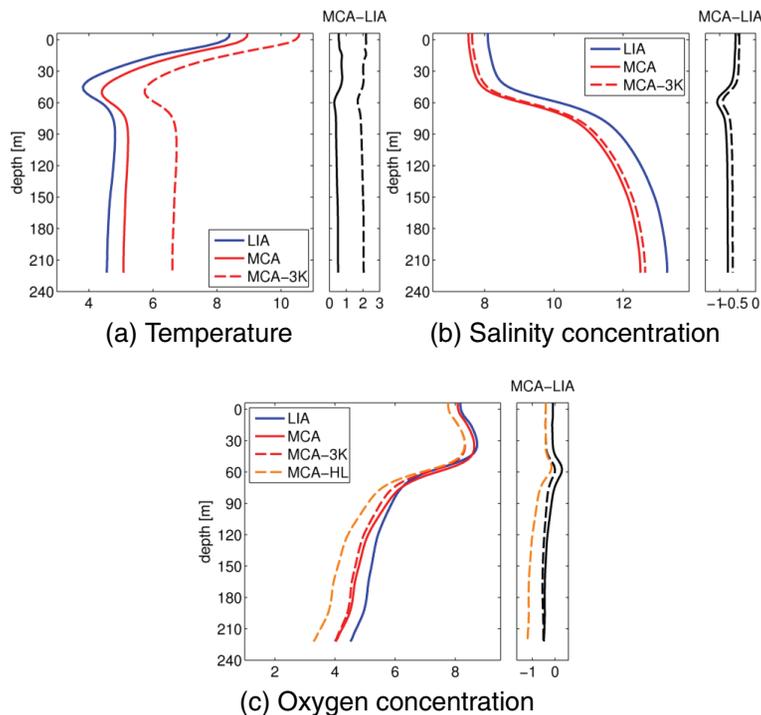


Fig. 9. Vertical profiles at Gotland Deep (location see Fig. 1). Solid lines show the profiles for RCO-MCA (red) and RCO-LIA (blue) while the dashed lines represent results from RCO-3K (red) and RCO-HL (orange), respectively. On the right sides, deviations from RCO-LIA are illustrated. Profiles are shown for temperature **(a)**, salinity **(b)**, and oxygen concentrations **(c)** and reflect means over the last 50 yr of the simulations.

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