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Volcanic impact on the Atlantic ocean over the last millennium

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Abstract

The oceanic response to volcanic eruptions over the last 1000 years is investigated with a focus on the North Atlantic Ocean, using a fully coupled AOGCM forced by a realistic time series of volcanic eruptions, total solar irradiance (TSI) and atmospheric greenhouse gases concentration. The model simulates little response to TSI variations but a strong and long-lasting thermal and dynamical oceanic adjustment to volcanic forcing, which is shown to be a function of the time period of the volcanic eruptions, probably due to their different seasonality. The thermal response consists of a fast tropical cooling due to the radiative forcing by the volcanic eruptions, followed by a penetration of this cooling in the subtropical ocean interior one to five years after the eruption, and propagation of the anomalies toward the high latitudes. The oceanic circulation first adjusts rapidly to low latitude anomalous wind stress induced by the strong cooling. The Atlantic Meridional Overturning Circulation (AMOC) shows a significant intensification 5 to 10 years after the eruptions of the period post-1400 AD, in response to anomalous atmospheric momentum forcing, and a slight weakening in the following decade. In response to the stronger eruptions occurring between 1100 and 1300, the AMOC shows no intensification and a stronger reduction after 10 years. This study thus stresses the diversity of AMOC response to volcanic eruptions in climate models and tentatively points to an important role of the seasonality of the eruptions.

1 Introduction

Understanding the climate fluctuations at decadal timescales and the climate response to external forcing is of prime importance to anticipate and understand future climate changes. The last millennium constitutes an interesting framework for investigating natural and forced variations, as climate reconstructions are reaching a relatively high temporal resolution (Jones et al., 2001; Mann et al., 2009), and show substantial decadal to multidecadal fluctuations (e.g. Gray et al., 2004). However, reliable oceanic

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reconstructions are still very rare (Sicre et al., 2008; Masse et al., 2008; Richter et al., 2009), mainly because of the difficulty to obtain undisturbed high sedimentation rate and well-dated marine sediments. Meanwhile, computer resources are increasing so that climate integrations using state-of-the-art coupled ocean-atmospheric general circulation models (OAGCM) are becoming routinely available over this period, allowing investigations of the mechanisms of low frequency climate variability.

Several observational studies have shown that the ocean, and in particular the North Atlantic, plays a large role in decadal climate variability (e.g. Knight et al., 2005; Sutton and Hodson, 2003, 2005). The relative importance of the various external forcings, however, remains debated. In model studies, this is partly due to their different representation and partly to divergent model responses. van der Schrier et al. (2002) and Hofer et al. (2011) suggested that external forcings, primarily variations of the total solar irradiance (TSI), act as modulators of the natural climate variability. Goosse and Renssen (2006) reported a decrease of the large scale Atlantic Meridional Overturning Circulation (AMOC) for increasing TSI, similar to the response to an increased atmospheric CO₂ concentration. In Zorita et al. (2004), on the contrary, the TSI does not have a significant impact on the AMOC.

Volcanic eruptions constitute another important external forcing over the last millennium. Their climatic impact has been largely investigated in terms of atmospheric thermal and dynamical anomalies, in relation with the North Atlantic Oscillation, the El Nino-Southern Oscillation or the monsoons systems (e.g. Oman, 2006; Shindell et al., 2004; Stenchikov et al., 2006; Trenberth and Dai, 2007). Probably because of the lack of reliable reconstructions, fewer studies have focused on their effect on the oceans. Church et al. (2005) and Gleckler et al. (2006) suggested that oceanic anomalies following a volcanic eruption could be lasting more than a decade. Using sensitivity response studies to relatively recent eruptions (the Pinatubo in 1991 and the Tambora in 1815), Stenchikov et al. (2009) showed that while radiative forcing produced by these explosive events lasted for about 3 years, the volcanically induced tropospheric temperature anomalies remained significant for seven years, the sea ice

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responded on decadal time scales, and the deep ocean temperature, sea level, salinity and the AMOC were perturbed for several decades to a century. In particular, the AMOC strengthened by roughly 10 %, but the amplitude to the response scaled less than linearly with the strength of the eruption. In response to a super-eruption with 100 time the Pinatubo amount of sulphuric acid released in the stratosphere, Jones et al. (2005) found that the AMOC doubled in intensity after nine years. Finally, both Ottera et al. (2010) and Ortega et al. (2011) found an intensification of the AMOC in a long simulation forced by both reconstructions of volcanic eruptions and variations in TSI. This response was associated to a persistent positive phase of the North Atlantic Oscillation (NAO) in Ottera et al. (2010) but not in Ortega et al. (2011).

Several studies have also begun to point out the specificity of the thirteenth century in terms of intense volcanic activity, and the possibility for a cumulative impact on the ocean (e.g. Zhong et al., 2010). The second half of the thirteen century is indeed the most perturbed half century of the past 1500 years (Jansen et al., 2007). In two out of four simulations, Zhong et al. (2010) found a centennial-scale climate change following the succession of decadal paced eruptions following the 1257–1258 mega-eruption. They highlighted a coupled ice-ocean interaction between the subpolar North Atlantic, a reduced extension of the AMOC into the northern North Atlantic, and the Arctic ocean, which maintained significantly expanded sea ice and reduced surface air temperatures for at least 100 years. However, the feedback mechanism depended on other factors since it was only activated in half of the simulations. In this context, it is important to highlight that while both Stenchikov et al. (2009) and Ottera et al. (2010) found a significant intensification of the AMOC following volcanic eruptions, the first study was based on sensitivity experiments following single eruptions of different intensities and the other based on composite analysis over the last 600 years of the millennium, thereby excluding the particular succession of events of the thirteen century.

Here, we explore the interannual to decadal oceanic response to volcanic activity in a coupled OAGCM forced by a full set of reconstructed external forcings over the last

millennium. Sicre et al. (2011) have shown that the simulated sea surface temperature (SST) in the northern North Atlantic compares well with a recent high resolution SST reconstruction off Iceland. We propose to describe more thoroughly the oceanic response to the major volcanic eruptions of the last millennium and investigate the mechanisms for the oceanic circulation adjustment. The model configuration and the forcings are presented in Sect. 2. The oceanic response to solar and volcanic forcing are compared in Sect. 3 and the temperature response to volcanic eruptions is discussed in Sect. 4. In Sect. 5, we investigate the response of the Atlantic circulation to isolated volcanic eruptions (occurring after year 1400) and in Sect. 6, we highlight the differences with the twelfth and thirteenth century. Conclusions are given in Sect. 7.

2 Model and experiment

2.1 The coupled model

We use the IPSLCM4 v2 climate model developed at the Institut Pierre-Simon Laplace (Marti et al., 2010). This model couples the LMDz4 atmosphere GCM (Hourdin et al., 2006) and the ORCHIDEE 1.9.1 module for continental surfaces (Krinner et al., 2005) to the OPA8.2 ocean model (Madec et al., 1998) and the LIM2 sea-ice model (Fichefet and Maqueda, 1997), using the OASIS coupler (Valcke et al., 2000). The resolution in the atmosphere is 3.75° in longitude, 2.5° in latitude, and 19 vertical levels. The ocean and sea-ice are implemented on the ORCA2 grid (averaged horizontal resolution $2 \times 2^\circ$, refined to 0.5° around the equator, 31 vertical levels). In all simulations, the vegetation was set to a modern climatology from Myneni et al. (1997). After a 310 year spin up with preindustrial greenhouse gases (GHG) concentrations and tropospheric aerosols, two simulations were run. The first one is a 1000-year control simulation (CTRL) with the same preindustrial conditions as the spin up, also used in Servonnat et al. (2010). The main characteristics of the AMOC in the model and its sensitivity to freshwater have been discussed by Swingedouw et al. (2007). They showed that

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an excess of freshwater flux over the Labrador Sea was responsible for the lack of deep convection in this region and the relatively weak AMOC (11 Sv) in the model. Deep convection in the northern North Atlantic only takes place in the Nordic Seas and south of Iceland (Marti et al., 2010). The natural variability of the AMOC, its link to deep convection and its impact on the atmosphere have been studied by Msadek and Frankignoul (2009). They showed that the multidecadal fluctuations of the AMOC are mostly driven by the deep convection in the subpolar gyre with a time lag of 6 to 7 years. Convection in the subpolar gyre is itself primarily influenced by anomalous salinity advection caused by the variability of the East Atlantic Pattern (EAP), second dominant mode of atmospheric variability in the North Atlantic region. The lack of Labrador Sea convection in the model probably explains the dominance of the EAP (as opposed to the North Atlantic Oscillation) in forcing multidecadal variations of the AMOC. The second simulation (LM2SV) was forced with a reconstruction of TSI, GHGs concentrations, changes in orbital parameters, and radiative effect of volcanic eruptions over the last millennium, from 850 to 2000 AD. The choice and implementation of the forcings are discussed below. To reduce the influence of the model drift, a quadratic trend was removed from each variable and grid point.

As our main focus is on the oceanic response to volcanic eruptions at interannual to decadal timescales, all data are considered in annual mean, or seasonal mean for such variables as sea ice cover and mixed layer depth.

2.2 External forcing over the last millennium

A number of different reconstructions for TSI variations have been produced (e.g. Jansen et al., 2007), mostly differing in the estimated reduction of total irradiance during the 17th century Maunder Minimum, which ranges from 0.08 % to 0.65 % (1.1 to 8.9 W m⁻²) of the contemporary value. Ammann et al. (2007) found that a TSI decrease of about 0.25 % during the Maunder Minimum produces a realistic amplitude of the Northern Hemisphere temperature change in climate models. However, recent progress in solar physics (Foukal et al., 2004; Solanki and Krivova, 2006; Gray et al.,

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2010) imply that the TSI variations between the Maunder Minimum and present day value are about 0.1 %. As this scaling is recommended for the third phase of the paleoclimate modelling inter-comparison project (PMIP III, Schmidt et al., 2011), we use the TSI reconstruction by Vieira and Solanki (2009) and Krivova et al. (personal communication, 2009), which follows it. The corresponding variations of the raw shortwave input at the top of the atmosphere is shown in Fig. 1 (top panel).

Large volcanic eruptions inject sulfur gases into the stratosphere, which convert to sulfate aerosols with a residence time of about a year. The aerosol cloud has several effects on radiative processes, most notably by backscattering part of the incoming solar radiation, which induces a net cooling at the Earth's surface (e.g. Robock, 2000). Thus, until recent years, most modelling groups (e.g. Jansen et al., 2007) have represented the volcanic forcing by altering the solar constant. Although such a coarse approach leads to hemispheric averages that compared reasonably well to a “blend” of proxy and/or instrumental reconstructions (e.g. Goosse et al., 2005; Stendel et al., 2006), it does not properly represent regional and seasonal variations. It is indeed known that the climatic impact of volcanic eruptions highly depends on the season and that latitudinal dependence of the cooling in the troposphere (warming in stratosphere) evolves for at least 2 to 3 years after the eruption. Furthermore, the volcanic aerosols serve as surfaces for heterogeneous chemical reactions that destroy stratospheric ozone, which controls solar energy absorption in the stratosphere. Its variations thus alter both the vertical temperature gradient between the troposphere and the stratosphere and the latitudinal temperature gradient in the stratosphere. We implemented in the IPSL model a new radiative module described in Khodri et al. (2011) that mimics the direct radiative effect of sulphate aerosols. The input time series is based on the monthly mean optical thickness latitudinal reconstruction by Ammann et al. (2003) and Gao et al. (2008) from 850 AD to present. The anomalous optical thickness is implemented in the tropical band between 20° S and 20° N, and transported poleward within 3 years according to a spreading function as in Gao et al. (2008). Figure 1 (second panel) illustrates the time series of implemented stratospheric volcanic aerosols optical depth. Note that

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a change in the global mean optical depth of 0.1 corresponds to a global anomalous radiative forcing of roughly -3 W m^{-2} . However, as discussed in Khodri et al. (2011) and Timmreck et al. (2009), the volcanic module tends to overestimate the radiative effect of the mega eruptions because of the use, for paleo-eruptions, of aerosol effective radius and optical depth derived from observations over the instrumental period, specifically for the Mount Pinatubo (1991) and El Chichón (1982) volcanic eruptions.

The greenhouse gas concentrations are those inferred from ice cores and direct measurements as reported in Servonnat et al. (2010). This simulation does not include the forcing by anthropic aerosols, so that global warming detected over the last decades of the simulation is overestimated (not shown). Hence, this study focuses on the natural external forcings and the period of investigation is limited to years 850 to 1849 AD.

3 Temperature response to solar and volcanic forcings

The most striking signal in the time evolution of the air temperature at 2m averaged over the Northern Hemisphere (Fig. 1 third panel) and the SST averaged over the Atlantic ocean (Fig. 1 fourth panel) are important variations following volcanic eruptions, in particular an abrupt cooling of up to 3°C in the atmosphere and 1°C in the ocean. Such signature has also recently been found in temperature reconstructions in the subpolar North Atlantic (Sicre et al., 2011). On the other hand, variations of the solar insolation do not seem to have a strong imprint. The lagged correlation r of the anomalous TSI time series with the averaged surface air temperature in the Northern Hemisphere and with the Atlantic SST have a broad but weak maximum when the TSI leads by 4 years, reaching $r = 0.12$ and 0.13 respectively (significant at the 5 % level) (Fig. 2, top panel). The corresponding correlation with the volcanic forcing is much larger, peaking when temperature lags by one year, with $r = -0.63$ and $r = -0.52$ respectively (Fig. 2, bottom panel). For both air and sea temperature, the correlation with the volcanic signal remains significant for more than 15 years. Note that the significant correlation

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at lag -1 in Fig. 2 is due to the use of annual averages, as eruptions might in fact have started during the calendar year preceding the maximum of emission. The stronger influence of volcanic forcing is probably due to our use of a TSI reconstruction with weak variations and to an overestimation of the volcanic radiative effect (Sect. 2.2).

The frequency dependence of the solar correlation is illustrated by the cross-wavelet coherence spectra in Fig. 3. The wavelet analysis was made with the Morlet wavelet, and the transform performed in Fourier space, using zero padding to reduce wraparound effects (Torrence and Compo, 1998). The parameters were chosen to give a total of 57 periods ranging from 0.5 to 256 years, and the square coherency were calculated using smoothing in the time and space domain (Grinsted et al., 2004), with the 5 % significance level determined from a Monte-Carlo simulation of 1000 sets of surrogate time series. The two temperature time series show episodic coherency with the solar forcing at 11 year period (Fig. 3, top panels), in particular around 1200 and 1600. Meehl et al. (2008, 2009) indeed showed that a peak in the solar activity induces surface cooling in the tropical Pacific. Kuroda et al. (2008) showed that over the historical period, years of anomalously high solar irradiance were associated with a large warming of the lower stratosphere through radiative heating. Such a temperature anomaly in the stratosphere creates anomalous temperature of opposite sign at lower heights. However, these processes require a much higher resolution in the stratosphere to be properly represented. In fact, episodic coherency between SST and TSI variations at 11 years timescale is also significant from the control data, suggesting that the signal in Fig. 3 is internal to the data sets and does not indicate physical response of the ocean to the 11-year cycle. The temperature time series also show strong coherency with the TSI variations at multidecadal timescale from 1700, associated to the TSI increase, and at centennial time scales over the whole simulation (Fig. 3, second and third panels).

It is somewhat more difficult to distinguish the response of the Atlantic meridional overturning circulation (AMOC) from its natural variability. As shown in Fig. 1 (bottom panels), the AMOC intensifies during the second half of the thirteenth century, when

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volcanic activity was intense, peaks around year 1280 and then rapidly decreases, reaching a minimum around year 1320, about 60 years after the major eruption of 1260. There is a hint of a weak response to the eruptive events in the early 1800s. As shown in Fig. 3 (bottom), there is a hint of a weak coherence between the time series of AMOC maximum with the TSI variations at 11-year periods, and a more significant one at about 100-year period, with TSI leading by 15 years. Correlations with the volcanic forcing are barely significant. As will be shown below, this does not imply that there is no AMOC response to natural forcings, in particular volcanic eruptions. Time series of AMOC maximum represents one mode of AMOC variability, namely a basin scale acceleration, as discussed for example in Msadek and Frankignoul (2009). More local AMOC adjustments require more specific analysis. In the following, we concentrate on the response to volcanic forcing, which has a much stronger impact on the atmospheric and the oceanic temperature than the solar forcing in the model.

4 Anomalous temperature patterns in response to volcanic eruptions

To describe the oceanic response to a volcanic eruption, we construct a composite evolution based on the oceanic anomalies that follow the major eruptions. Anomalies are computed for each selected eruption as the difference between the time evolution of the field and a reference defined as the average of the field during the two years preceding the eruption. Composites are then defined as the average of these anomalies scaled by the magnitude of each eruption, so that possible non linear effects linked to the eruption magnitude are minimized. Note however that our conclusions are unchanged without this normalization. In order to maximize the signal to noise ratio, we focused on relatively large eruptions, and thus selected eruptions corresponding to an increase of stratospheric aerosol optical depth (AOD) by more than 0.15 (eruptions marked with a star in Fig. 1), which is equivalent to a global radiative forcing of at least -2.8 W m^{-2} . This corresponds to the 9 strongest eruptions between 850 AD and 1849. As seen in Fig. 1 and discussed in Sect. 2.2, several of the selected events follow each other by

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less than 10 years (1169–1178, 1810–1816). To minimize the interference between successive events, eruptions which precede another one by less than the considered time lag in the composite were omitted. As a consequence, the number of events in the composites may decrease with lag. The composites are displayed for a stratospheric global mean optical depth equal to 0.15. Significativity is tested with a block bootstrap procedure with 500 permutations of the volcanic time series in blocks of 3 years (the maximum residence time of stratospheric aerosols).

Figure 4 shows composites of anomalous global surface temperature up to 20 years after a volcanic eruption of AOD of 0.15. The first panel (year 0) shows that the maximum cooling occurs in the tropics and on the lands during the year of the eruption. There is also a meridional dipole in the Southern Atlantic and the Indian oceans, which can be shown to be due to a shift of the westerlies, persisting for a year. An anomalous warming in the polar region over Eurasia is consistent with the observations (e.g. Robock and Mao, 1992) and closely related to tropospheric and stratospheric circulation changes. As the lag increases, the tropical oceanic signal extends in latitude, reflecting the spreading of the atmospheric cooling (e.g. Robock, 2000), while progressively decaying in the tropics. In the subpolar North Atlantic, the cooling peaks at year 3 and decays thereafter. Note the relatively rapid decay of the cooling in the eastern equatorial Pacific at year 1, also present at year 2 (not shown) which could be due to an El Niño-like response. One to two years after the eruption, there is an anomalous warming in the North Atlantic midlatitudes, the origin of which is discussed below. An anomalous warming near the Drake passage becomes significant at year 3 and reaches its maximum at year 5. Ten years after the eruption, the whole tropical band is still significantly anomalously cold, as well as some land areas such as in Eurasia. In the North Atlantic, the most striking feature is an anomalous warming in the Labrador Sea, which decays thereafter.

Figure 5 shows similar composites for the zonally averaged global oceanic temperature response as a function of depth up to 20 years after a volcanic eruption. Consistent with Fig. 4, a temperature decrease of up to 0.25 K appears in the upper tropical ocean

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during the eruption year, together with a warming below 100 m depth in the deep tropics. The latter results from a thickening of the tropical thermocline and a decrease of equatorial ventilation following a weakening of the trade winds, as discussed below. The surface cooling already reaches 60° N, but its poleward extension is stronger one year after the eruption, consistent with Fig. 4. By year 1, the signal has penetrated in the ocean interior around 30° N and 30° S, where oceanic ventilation mostly takes place. In the subtropics, the downwelling is shifted slightly poleward of the climatological ventilation region, indicated by the mean isotherms in Fig. 5 (white contours). As in Laurian et al. (2009), the shift can be explained by the poleward displacement of the surface isopycnals resulting from the surface cooling. Deep penetration down to 900 m of the cooling is also seen around 60° N at year 1, reflecting enhanced deep convection. Deep convection also mixes the cooling signal down in the Southern Ocean, reaching its largest depth 2 to 3 years after the eruption (not shown). In the following years, the tropical surface cooling decays, while persisting at depth and deepening further (Fig. 5, year 5). As the surface cooling reaches greater depths, the subsurface warming deepens and shifts poleward.

After 10 years, the cooling signal has reached more than 500 m at 40° latitude north and south, which is roughly the maximum depth of the subtropical cells, and 700 m in the southern ocean. In the North Atlantic, on the other hand, a warm subsurface anomaly has appeared, reflecting a decrease of deep convection as will be discussed below. At this stage, the response is thus asymmetric in the high latitudes as also found by Stenchikov et al. (2009). Twenty years after an eruption, cooling is still significant in the tropics and at high latitudes, where it reaches 700 to 900 m, while the northern subtropics have warmed, reflecting the dynamical adjustment of the gyres discussed below.

In the following, we focus on the response of the Atlantic Ocean, as a case study and in order to investigate the behavior of the AMOC. From Fig. 1, it seems clear that the behavior of the AMOC after the severe and decadal paced eruptions of the twelfth and thirteenth century is peculiar. Figure 6 illustrates the different response of the ocean

to the selected eruptions occurring after 1400, from the ones occurring between 1100 and 1300. In response to volcanic eruptions occurring after 1400 (bottom panels), the initial (in phase) cooling is more clearly limited to the tropics and subtropics, while the mid- and high latitudes are characterized by an anomalous warming, due to anomalous turbulent heat fluxes as discussed below. The anomalous cooling rapidly reaches the higher latitudes (1 yr to 4), except for a small patch of anomalous warming at 45° N which reflects a northward shift of the North Atlantic Current. After about a decade, the anomalous cooling has disappeared or lost significance in the Atlantic basin, while a strong and persistent warming has appeared in the subpolar gyre, with maximum amplitude south of Greenland, and a comma shape extension in the eastern subtropics with resembles the path of the subtropical gyre. This structure thus is strongly similar to the signature of an AMOC acceleration in the coupled model (e.g. Msadek and Frankignoul, 2009).

On the other hand, the anomalous cooling occurring in phase with the intense and decadal paced eruptions between 1100 and 1400 is significant not only in the tropics, but also at subpolar latitudes, in particular in the Irminger Sea and the Nordic Seas, where deep convection in the model takes place. There is also a weak, marginally significant, warming at midlatitude, again probably reflecting a shift in the North Atlantic current, but it is short lived and the entire basin becomes anomalously cold in the years following the eruption. At decadal timescales, the anomalous subpolar warming seen after 1400 can be recognized but it is much weaker and not significant at the 5 % level. The fact that the response response differs as early as in phase with the eruption tends to eliminate the cumulative effect of the decadal paced eruptions of the twelfth and thirteenth century, as opposed to more isolated eruptions occurring after year 1400. A larger signal to noise ratio in response to stronger eruptions might be an alternative explanation, as discussed e.g. in Shindell et al. (2003) and Schneider et al. (2009). However, the anomalous atmospheric response shown in Fig. 6 (top) is unchanged if the mega eruption of 1258–1259 is omitted for the computation of the composite (not shown). Table 1 suggests rather that the eruptions of the middle age period tend

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to peak during the cold season while the ones that occurred during the rest of the last millennium mostly peak during the warm season. Investigating the effect of this seasonality requires specific sensitivity experiments and is beyond the point of this study. In the following, we will first investigate the response to eruptions occurring after 1400.

5 Interannual to decadal response of the Atlantic ocean to eruptions post 1400 AD

In response to the rapid surface cooling, there is a strong anomalous low over the Canadian archipelago and an anomalous high over the northeastern Atlantic. In addition, the sea level pressure (SLP) becomes anomalously high over South America and most of Africa, where the cooling is strongest, and an anomalous low in the western subtropics (Fig. 7). Consequently, the Northern Hemisphere trades and westerlies are reduced during the year of the eruption, and shifted southward. Over the tropical lands, the SLP signal weakens at year 1, but it remains significant for almost 2 decades over the amazonian basin. At mid to high latitudes, the anomalous low quickly disappears but the anomalous anticyclone shifts slightly westward and persists until year 4, resembling a negative phase of the East Atlantic Pattern (EAP). Later, the signal loses significance (not shown), until year 10, where a response resembling a negative phase of the NAO is detected.

At year 0, the wind changes induce a negative wind stress curl anomaly across the basin between 50 and 60° N and a positive one north and south of it (Fig. 8, left). The depth-integrated oceanic circulation, as described by the barotropic streamfunction, adjusts rapidly to the wind stress curl. At year 0, it is anomalously negative in much of the subtropical Atlantic, reflecting a weakening of the subtropical gyres (Fig. 9, left). A weak positive anomaly is also significant in the subpolar region, where the wind stress curl is negative. At following lags, the negative wind stress curl anomaly persists at subpolar latitudes and shifts to the southern Irminger Sea, consistently with the SLP

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response (Fig. 8, right). Two to four years after the eruption, both gyres of the North Atlantic are thus clearly reduced. At longer lags, the response decays in the subtropics while the subpolar gyre stays anomalously weak for more than a decade after the eruption (Fig. 9, middle). Note also the persistent signal in the Labrador Sea where the cyclonic circulation is reinforced.

The atmospheric response to the eruption also induces vertical circulation in the ocean, resulting from the anomalous Ekman suction at 30° N/S and pumping around 50° N. This appears clearly at year 0 on the meridional streamfunction composite (Fig. 10, left). The signal is equivalent barotropic, with an upwelling around 30° N and a downwelling at 50° N and around the equator. During the following years, the strong negative wind stress curl in the subpolar North Atlantic maintains a positive meridional cell between 20 and 50° N, which can be viewed as an intensification of the AMOC in the North Atlantic, consistent with previous studies (e.g. Stenchikov et al., 2009; Ottera et al., 2010; Ortega et al., 2011). Note that this positive anomaly is probably also favored by the intensified deep convection that occurs during the year of the eruption (Fig. 11, left), and is associated with strong surface cooling. An intensification of deep convection typically leads by several years an acceleration of the AMOC in the North Atlantic basin (e.g. Mignot and Frankignoul, 2005). However, it is short-lived here, losing significance by year 1, so that the AMOC intensification does not persist more than a few years (Fig. 10, bottom left). On the other hand, there is a weak reduction of the AMOC north of about 60° N up to four years after an eruption, which later intensifies and extends to subpolar latitudes as a result of a reduction in deep water formation, as discussed below.

Five years after the volcanic eruption, the SLP anomaly decreases (not shown). Nevertheless, as indicated above, a significant SLP anomaly appears again near year 10–12, under the form of the dipole in the mid to high latitudes bearing similarity with a negative phase of the NAO. This could reflect the SLP response to the AMOC intensification seen at year 2–4 (Fig. 7 upper right), since Gastineau and Frankignoul (2011) found a weak but significant response of the atmosphere (negative NAO phase)

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to enhanced AMOC in several climate models including the CTRL simulation with IPSLCM4. In the latter, the SLP response was of similar magnitude and most clearly seen four years after an AMOC intensification. Here, the AMOC intensification indeed remains significant until lag 8 (not shown) Whether there is a link with the weak AMOC intensification seen about 20 years after the eruption cannot be asserted here but could be established in dedicated experiments.

As mentioned above, the strong surface cooling rapidly deepens the mixed layer south of Iceland (Fig. 11, left panel) and in the subtropics (not shown). The intensification of deep convection favors the penetration of the cooling signal at depth seen in Fig. 5 at high northern latitudes and a weakly significant retreat of sea ice cover (Fig. 12, left panel). However, this response loses significance in the following years (Fig. 11, middle panel), and instead, deep convection is reduced both in the Nordic Seas and South of Iceland 4 years after an eruption (Fig. 11, left panel). This persists for about a decade after the eruption. In the Nordic Seas, the reduction of deep convection is due to a persistent sea ice capping of the area during winter, resulting from the strong surface cooling (Fig. 12). This anomaly appears about 1 year after the eruption, peaks after four years and, again, persists for roughly a decade. Such anomalous sea ice extension is consistent with the sea ice reconstruction off Iceland from Masse et al. (2008), showing abrupt events that coincide with the volcanic eruptions after 1300 AD.

South of Iceland, the winter mixed layer shallowing in Fig. 11 (right) is due to a strong negative salinity anomaly (Fig. 13 top). The latter is largely due to anomalous Ekman transport (Fig. 13, bottom), while anomalous atmospheric freshwater fluxes play a lesser role (Fig. 13, middle), consistent with Mignot and Frankignoul (2003, 2004). Note that the anomalous surface freshwater flux caused by the volcanic eruptions are maximum during the year of the eruptions, and in the tropics (Fig. 13), consistent with Trenberth and Dai (2007). The reduction of the northern trade winds (Fig. 7) is indeed associated to a northward shift in the inter-tropical convergence zone. As a result, precipitation is enhanced near 10° N and strongly reduced along the equator, including over the continents. In the northern deep tropics, evaporation is reduced, again

because of reduced winds and SST. These anomalous surface freshwater fluxes induce the negative salinity anomaly in the northern subtropics in the years following the eruption.

6 Interannual to decadal response of the Atlantic ocean to eruptions between 1100 and 1300

Figure 14 shows the response of the AMOC to the volcanic eruptions selected during the period of intense volcanic activity between 1100 and 1300. As during the later period, the in-phase response is essentially characterized by an anomalous downwelling around 30° N. The associated negative and positive cells south and north of this latitude have nevertheless a much weaker extension in depth (for the tropical one) and in latitude (for the northern one). Indeed, the anomalous sea level pressure induced during the year of an eruption occurring during the earlier period, shown in polar view in Fig. 15, is similar in the tropics and subtropics (not shown) but it has the opposite sign over the canadian archipelago and there is no strong high in the eastern North Atlantic. As a result, the anomalous wind stress curl is much weaker in the subpolar region, inducing a much weaker anomalous Ekman pumping and Ekman transport, and only little salt advection (not shown). The subpolar gyre is thus much less affected during the years following the eruption (not shown). Lacking the large subpolar freshening seen in the later period, the winter mixed layer remains anomalously deep in the subpolar region (Fig. 16, top row middle panel), even further south than deep convection locations, due to the surface cooling (Fig. 6), which might contribute to the broad and shallow AMOC intensification between 20° S and 45° N at years 2 to 4. It can be shown that the tropical part of this anomalous cell is associated with an equatorward shift of the subtropical gyre. On the other hand, deep convection is reduced in the Nordic Seas after a few years because of the sea ice capping discussed above. Note that in this period, the anomalous sea ice extension establishes faster and it is stronger and more persistent (Fig. 16, bottom) than in the later period, consistent with

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the stronger cooling (Fig. 6). It is also interesting to note that the weak gyre response also contributes to maintain the oceanic surface cooling in the 1100–1300 A.D. period, since it does not induce the warm anomaly seen in the years following an eruption occurring after 1400 AD (Fig. 6). As a result, the negative AMOC anomaly seen over the ridges two to four years after an eruption is stronger and deeper than for the eruptions occurring after 1400 AD. After a longer delay, when anomalous cooling and convection in the subpolar basin start to vanish, the shallowing of the winter mixed layer in the Nordic Seas seen at years 4–7 in Fig. 16 (top right) persists and is likely responsible for the stronger negative AMOC weakening seen throughout the northern North Atlantic at least two decades after the eruptions (Fig. 14, bottom left). This behavior is consistent with Zhong et al. (2010). That the AMOC weakening extends north of the ridges adds credit to the implied role of Nordic Seas convection and could explain why the AMOC behavior has become nearly opposite to that seen after 1400 AD.

7 Conclusions and discussion

In this study, we have investigated the oceanic response to volcanic eruptions over the last thousand years, with a focus on the North Atlantic Ocean. We used a fully coupled AOGCM forced by a realistic chronology of volcanic eruptions, variations of the TSI and of the atmospheric greenhouse gases concentrations. The analysis highlighted the multiple timescales of the response, including a fast tropical temperature adjustment to the strong volcanic-induced radiative forcing, dynamical adjustment in response to the associated atmospheric circulation modifications persisting roughly 5 years, and a subsequent adjustment of the AMOC in response to anomalous convection at high latitudes. The analysis also highlighted differences in the response during two distinct periods of the last millennium.

The global surface temperature response is maximum one to two years after a volcanic eruption. The anomaly penetrates at depth via subtropical oceanic ventilation as

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well as deep convection at high latitudes. It thus persists globally in the ocean for more than 20 years. During the year of the eruption, anomalous tropical cooling induces an anomalous high over the continents and a reduction of the trades in the Atlantic ocean. The atmospheric response at mid to high latitudes depends on the eruptions. In this study, in order to investigate apparent discrepancy found in the literature regarding the AMOC response, we only separated eruptions occurring after 1400 from the ones occurring between 1100 and 1300. In the later period, the anomalous atmospheric structure in response to an eruption induces strong wind stress curl anomalies over the North Atlantic ocean, which lead to an important dynamical adjustment in the Atlantic ocean at interannual timescales, namely one to five years after an eruption. This adjustment of the oceanic circulation is equivalent barotropic, with an upwelling around 30° N and a downwelling at 50° N and around the equator. The anomalous vertical oceanic circulation can reach the full depth of the ocean. During the following years, the atmospheric structure evolves inducing an anomalous acceleration of the AMOC in the subpolar basin. However, this anomaly does not persist more than a few years, because of reduced deep convection in the high northern latitudes under the effect of anomalous sea ice extension and surface freshening that develop a few years after the eruption. A weak reduction of the AMOC is thus detected a decade after the eruption.

In the case of the eruptions occurring between 1100 and 1300, the anomalous SLP structure during the year of the eruption differs over the subpolar region from the one obtained during the later period. It induces much weaker wind stress curl anomalies over the Atlantic basin and thus a much weaker dynamical adjustment of the AMOC in the years following an eruption. On the other hand, the initial reduction of deep water formation is more persistent, as a result of a stronger surface cooling and more persistent sea ice cover anomalies, leading to a stronger negative anomaly of AMOC at high latitudes 2 to 4 years after an eruption, and a stronger reduction of the AMOC in the subpolar North Atlantic 10 to 15 years after the eruption.

As noted in the introduction, the oceanic response to volcanic eruptions is still largely unknown and recent studies based on climate models suggest either an AMOC

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enhancement or a reduction following volcanic eruptions of the last millennium. The present findings could possibly reconcile these previous studies, suggesting a strong sensitivity of the response to different volcanic eruptions. In particular, the AMOC intensification seen 5 to 10 years after the volcanic eruptions occurring after 1400 AD. bears strong similarity with results of Ottera et al. (2010) using a different coupled climate model to investigate this period. On the other hand, the AMOC weakening in the northern North Atlantic and the large sea ice extension following the intense eruptions occurring between 1100 and 1300 can be compared to the response found by Zhong et al. (2010). The present analysis suggests that these different responses involve in fact similar mechanisms, namely an initial dynamical adjustment to anomalous winds and a subsequent thermohaline response to anomalous deep convection. However, the atmospheric response to the volcanic eruptions differ in the two periods and thus plays a large role in modulating the oceanic response. At least three factors could explain why the atmospheric response seems to change in time: the seasonality of the eruption, its intensity, and the cumulative effect in the case of successive eruptions. An analysis of these various effects requires specific experiments that are left for future studies. However, the results presented here suggest that the seasonality is the most plausible explanation. Indeed, the eruptions during 1100–1300 occurred mostly during the cold season while those after 1400 occurred mostly during the warm season. Although cumulative effects may also play a role, they cannot explain why the short term response already differs in the two periods. Non linearities in the response could also be important, even though our results were not changed when the mega eruption of 1258 was omitted from the 100–1300 composite.

The oceanic response is also likely to be affected by model biases and experiment design. In particular, the lack of deep convection in the Labrador Sea in the IPSLCM4 model is probably a major drawback. A better representation of the stratosphere is also needed to improve the representation of the effect of volcanic aerosols. Indeed, stratospheric dynamics and chemistry may significantly alter the modeled climatic impact of volcanic eruptions. This should be tested in the new version of the IPSL model

including 39 atmospheric levels and an improved radiative module, currently under development, as well as in the forthcoming CMIP5 database.

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Table 1. Month of maximum global mean AOD for the selected eruptions. The input time series is based on the monthly mean optical thickness latitudinal reconstruction by Ammann et al. (2003) and Gao et al. (2008).

Period 1100–1300		Period 1400–1850	
eruption year	month of maximum AOD	eruption year	month of maximum AOD
1168	February	1452	May
1177	February	1641	May
1231	November	1809	May
1258	April	1815	September
1257	May to October		

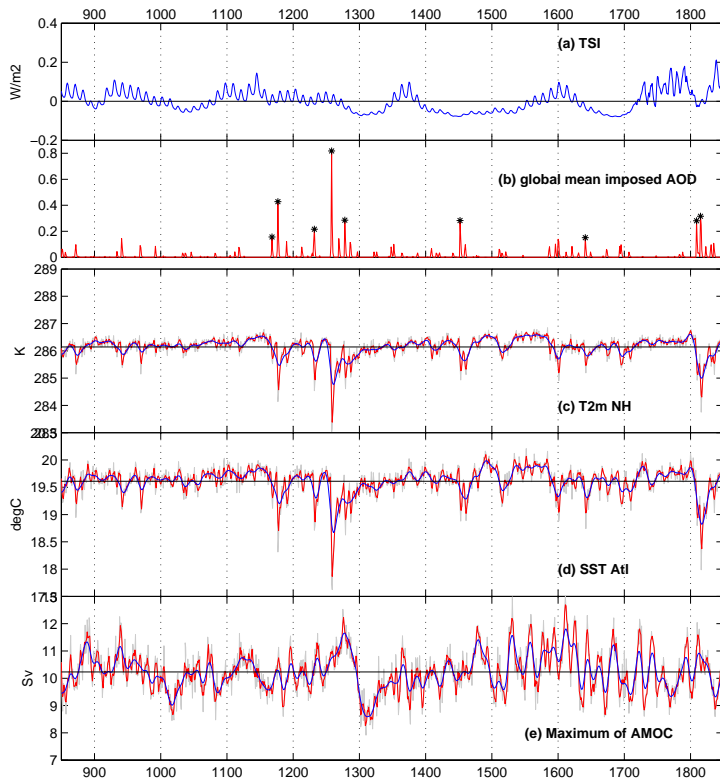


Fig. 1. Time series from 850 to 1850 of **(a)** anomalous short wave input at the top of the atmosphere, taken as an estimation of variations of the TSI, **(b)** imposed optical depth of volcanic aerosols. **(c)** Northern Hemisphere air temperature at 2 m **(d)** SST averaged between 30° S and 70° N in the Atlantic **(e)**. Maximum of the meridional overturning circulation between 10° N and 60° N and below 500 m depth in the Atlantic. For **(c)–(e)** raw annual mean data are shown in grey and low pass filtered data using a running mean of 3 years in red. Low pass filtered data using a spline function with a cutoff at 20 years are shown in blue.

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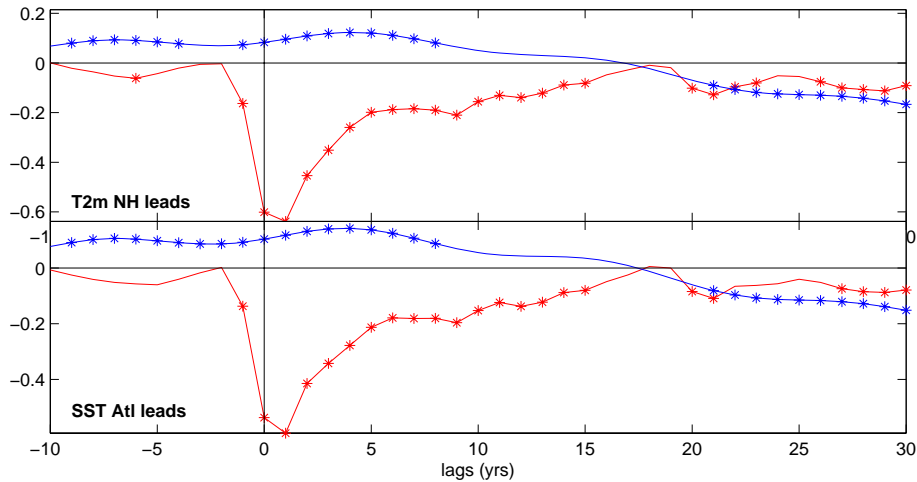


Fig. 2. Cross correlation between the temperature time series in Fig. 1 and the time series of the solar forcing (blue) and the volcanic forcing (red) from 850 to 1850 AD. Lags with a star are significant at the 5% level, tested against a bootstrap procedure with 500 permutation of the forcing time series using blocks of 3 years.

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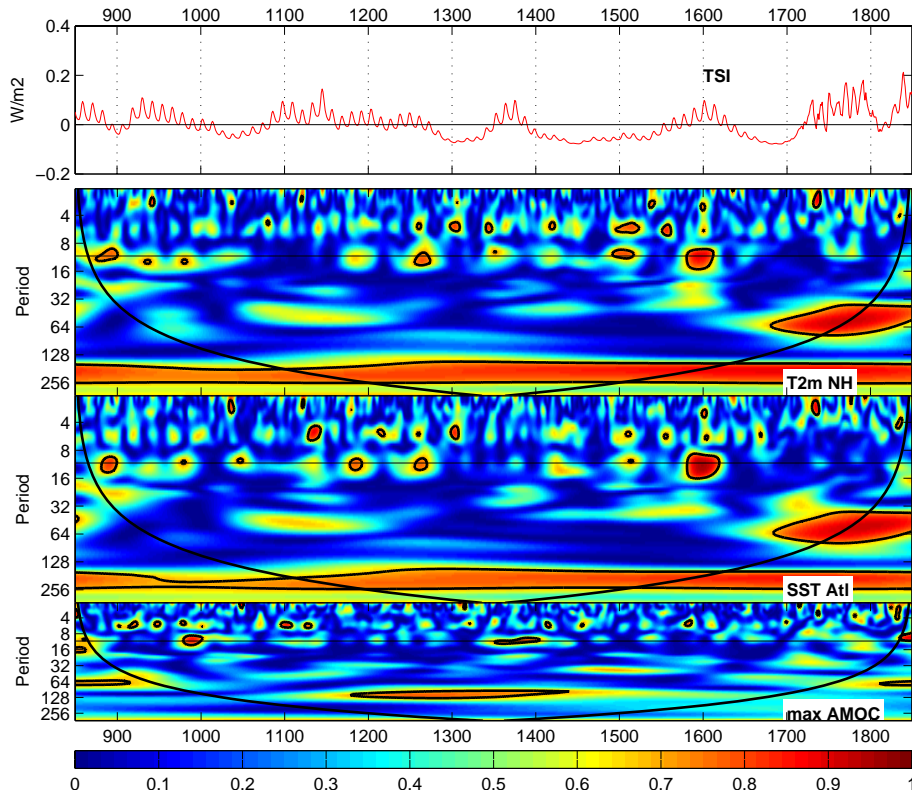


Fig. 3. Time series of the solar forcing (top) and cross-wavelet coherence spectrum of the temperature time series in Fig. 1 (bottom three panels). The thick contours enclose regions of greater than 95 % confidence and the thin lines indicate the limit of the cone of influence. The horizontal line indicates the 11 year period.

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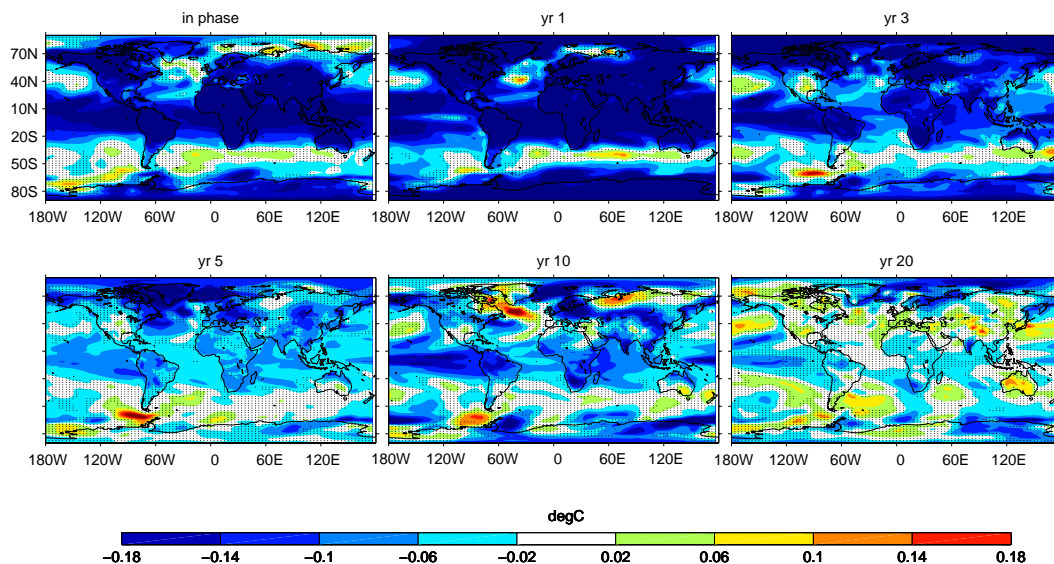


Fig. 4. Composites of anomalous surface temperature for different time lags (in years) with respect to the volcanic eruptions corresponding to optical depths higher than 0.15. Surface temperature corresponds to SST over the ocean and air temperature otherwise. Dotted areas are not significant at the 5 % level.

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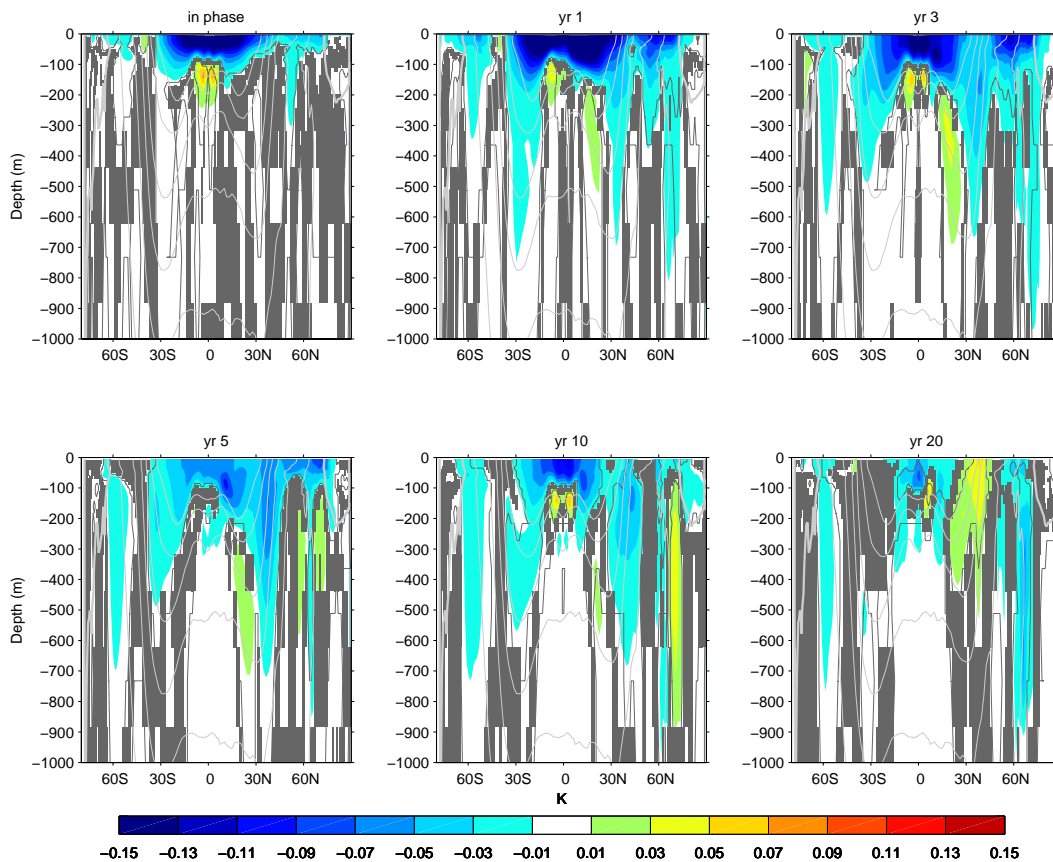


Fig. 5. Composites of anomalous zonally averaged oceanic temperature response at different time lags (in years). Shaded areas are not significant at the 20 % level. Black contours indicate anomalies significant at the 5 % level. Grey contours show the zonal mean global temperature (contour interval is 3 °C).

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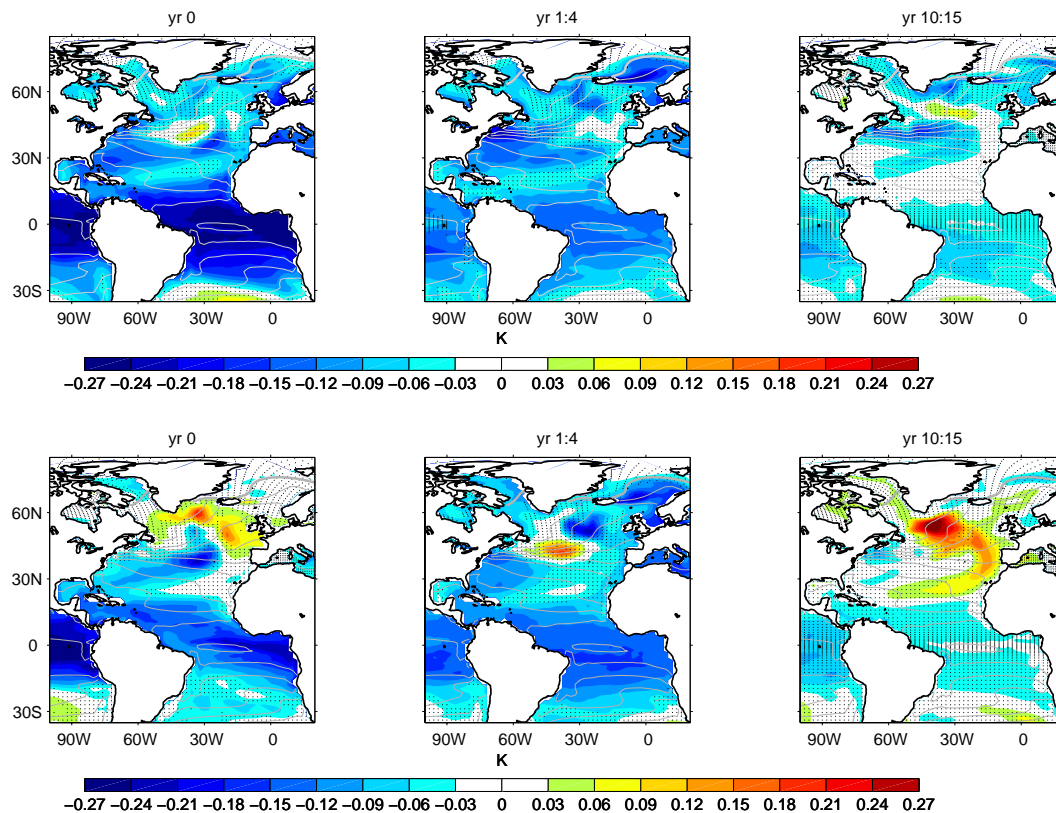


Fig. 6. Composites of anomalous sea surface temperature following volcanic eruptions in different periods. Top: between 1100 AD and 1300 AD, bottom: after 1400 AD. Dotted areas are not significant at the 5% level. Grey contours show the annual mean SST. Contour interval is 3 K, the thick line is for the zero contour.

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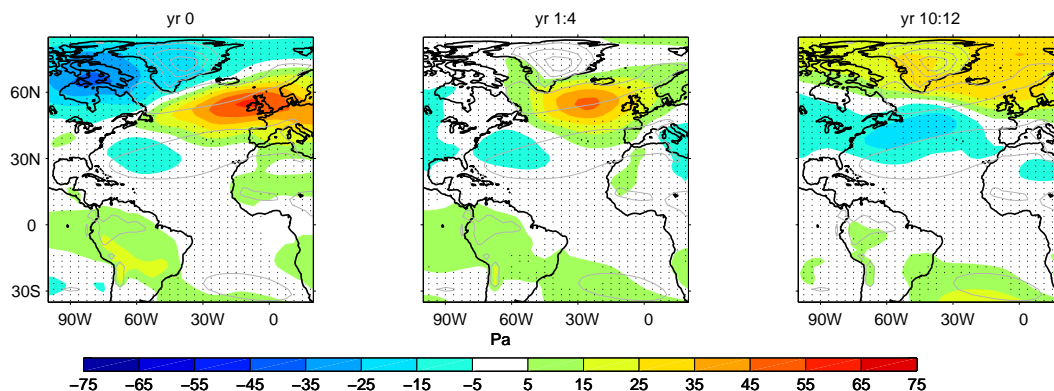


Fig. 7. Composites of annual mean anomalous sea level pressure in phase with the volcanic eruption (left) and following the volcanic eruption (other panels) for the period after 1400 AD. Grey contours show the SLP annual mean in the model (contour interval is 10^2 Pa).

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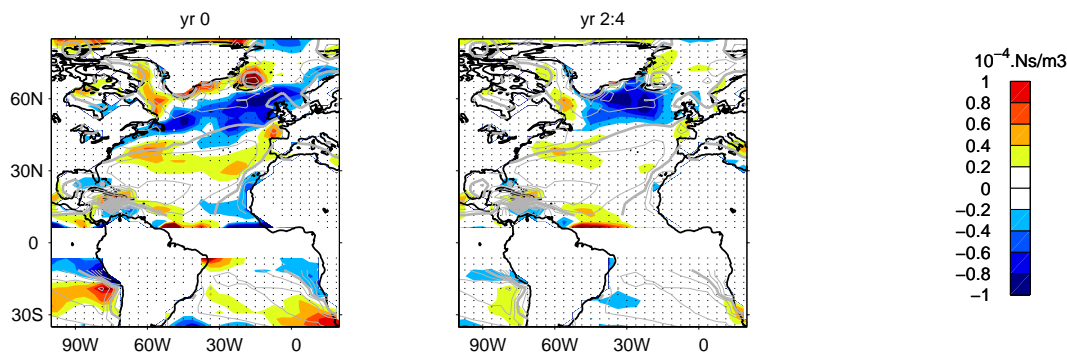


Fig. 8. Composites of anomalous wind stress curl, shown over oceans only, in phase with the volcanic eruption (left) and following the volcanic eruption by two to four years (right) for the period after 1400 AD. Grey contours show the annual mean values (contour interval is $10^{-4} \text{ N s m}^{-3}$), the thick grey line shows the zero contour.

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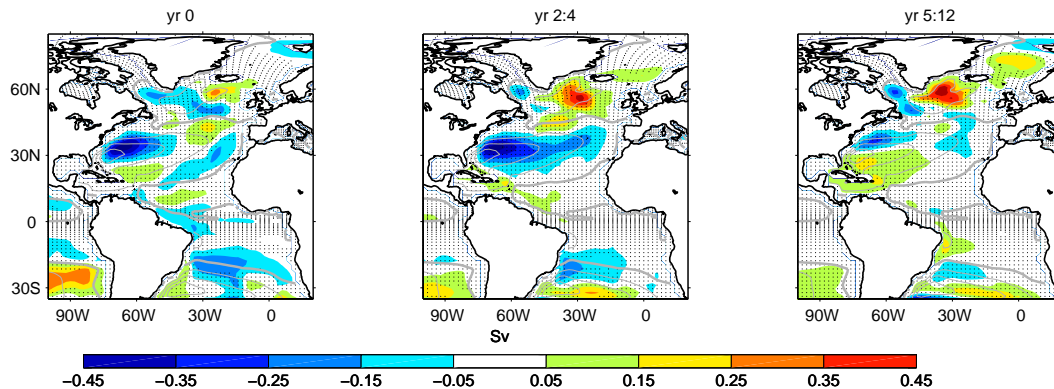


Fig. 9. Composites of anomalous Atlantic barotropic streamfunction in phase with the volcanic eruption (left) and following the volcanic eruption (other panels) for the period after 1400 AD. Positive (negative) values correspond to an anticyclonic (cyclonic) circulation. Grey lines show the annual mean SSS field in the model, with a contour interval of 10 Sv.

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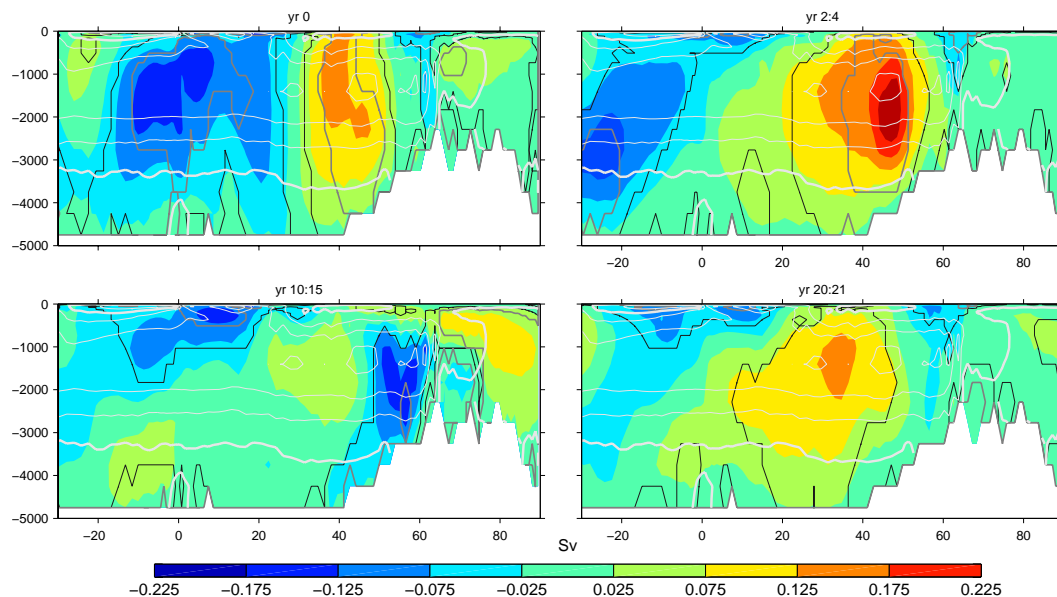


Fig. 10. Composites of anomalous Atlantic meridional streamfunction in phase with the volcanic eruption (left) and following the volcanic eruption (other panels) for the period after 1400 AD. Positive (negative) values correspond to a clockwise (counter-clockwise) circulation. Shading mask non significant areas at the 20 % level according to the Monte Carlo permutation test and black lines mark significant areas at the 95 % level according to the same test. Grey contours show the annual mean Atlantic meridional circulation in the control simulation (contour interval is 3 Sv, thick contour corresponding to the zero contour.)

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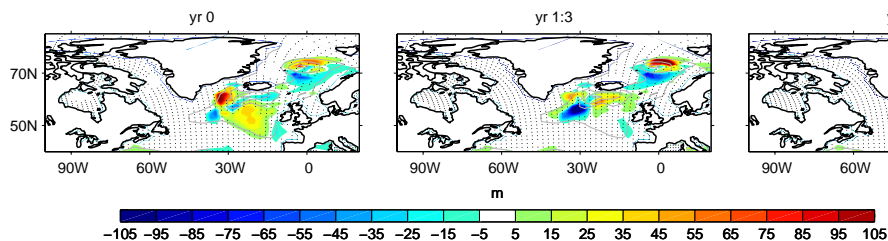


Fig. 11. Composites of anomalous March mixed layer depth during the year of the eruption (left), averaged over the following 3 years (middle), and averaged 4 to 7 years later (right) for the period after 1400 AD. Grey lines show the annual mean mixed layer depth field in the model, with a contour interval of 500 m.

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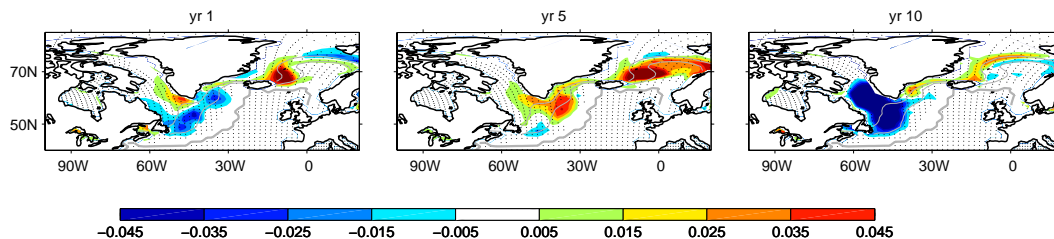


Fig. 12. composite of anomalous March sea ice cover during the year following the eruption (left), 5 years later (middle), and 10 years later (right) for the period after 1400 AD. Grey lines show contours of the average sea ice cover in March of 0, 0.5 and 1.

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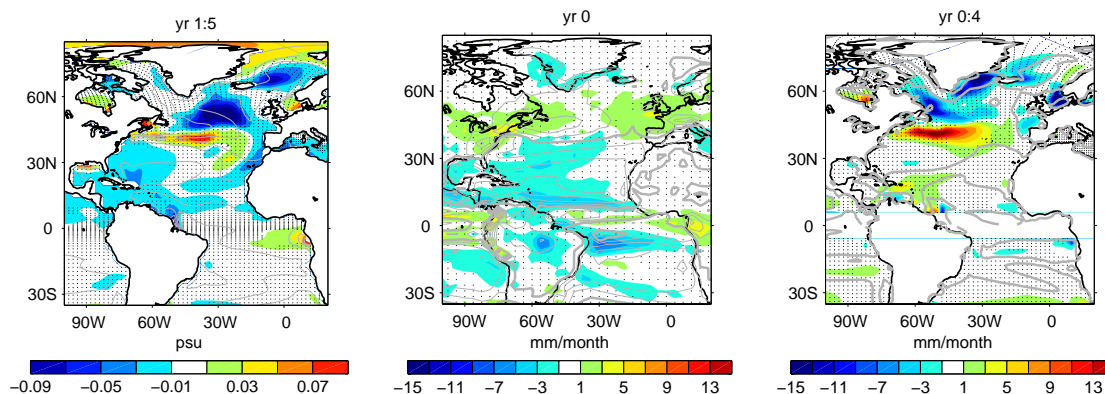


Fig. 13. Left: composites of anomalous SSS averaged 1 to 5 years after the eruption. Grey lines show the annual mean SSS field in the model, with a contour interval of 2 psu. Middle: composite of anomalous E-P during the year of the eruption. Right: composite of advection of mean salinity by anomalous Ekman currents (with a negative sign in order to be consistent with the E-P forcing term) averaged over the 4 years following the eruption. Values equatorward of 10. latitude were are masked because undefined. All composite are computed for eruptions occurring later than 1400 AD. For the two right panels, grey lines show the annual mean field in the model with a contour interval of 40 mm month⁻¹. The zero contour is thicker. For all panels, dotted areas are not significant at the 5 % level.

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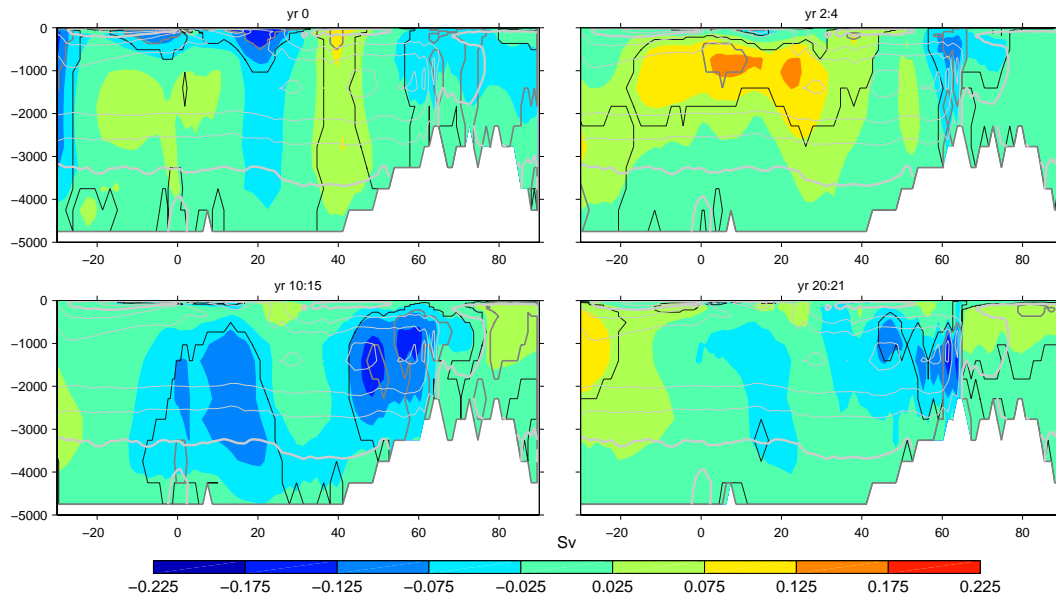


Fig. 14. As in Fig. 10 for the Atlantic meridional streamfunction during the period 1100–1300. Positive (negative) values correspond to a clockwise (counter-clockwise) circulation. Shading mask non significant areas at the 20 % level according to the Monte Carlo permutation test and black lines mark significant areas at the 95 % level according to the same test. Grey contours show the annual mean Atlantic meridional circulation in the control simulation (contour interval is 3 Sv, thick contour corresponding to the zero contour).

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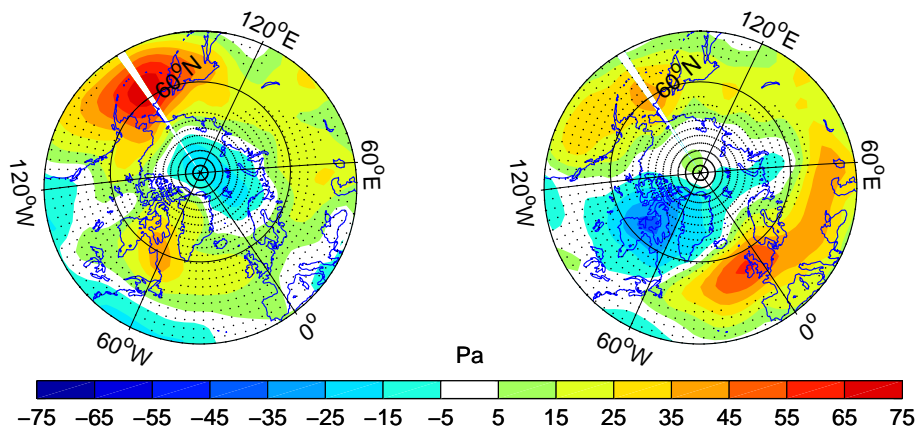


Fig. 15. Composite of annual mean anomalous sea level pressure in phase with the volcanic eruptions of the period 1100–1300 AD (left) and with the volcanic eruptions post-1400 AD (right).

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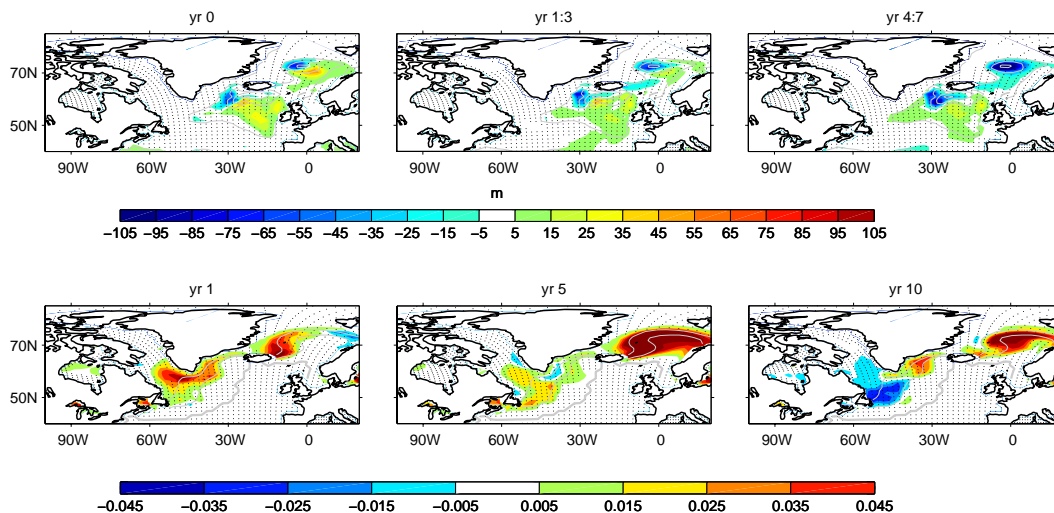


Fig. 16. Top: composites of anomalous March mixed layer depth in response to volcanic eruptions occurring during the period 1100–1300. The composite is shown for the year of the eruption (left), averaged over the following 3 years (middle), and averaged 4 to 7 years later (right). Grey lines show the annual mean mixed layer depth field in the model, with a contour interval of 500 m. Bottom: composites of anomalous March sea ice cover in response to volcanic eruptions occurring during the period 1100–1300. The composite is shown for the year following the eruption (left), 5 years later (middle), and 10 years later (right). Grey lines show contours of the average sea ice cover in March of 0, 0.5 and 1. Dotted areas are not significant at the 5 % level.

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