Impacts of Atlantic multi-decadal variability on the Indo-Pacific and Northern Hemisphere climate

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Abstract

Earlier studies have shown that Atlantic multi-decadal variability (AMV) can impact climate variability globally. However the instrumental records are short compared to the timescale of AMV, and mechanisms for these impacts are unresolved. This thesis deals with impacts of AMV on the Indo-Pacific region and over the Northern Hemisphere, investigating both the persistence of multi-decadal variability in the two phenomena and possible mechanisms for interactions between them. Coupled climate models are the main tools used in this study, in the form of output from the Coupled Model Intercomparison Project 5 (CMIP5) ensembles, a freshwater hosing experiment with the Bergen Climate Model (BCM), and partially coupled ensemble simulations with the Norwegian Earth System Model (NorESM). In addition proxy records are used to assess persistence in both AMV and its relation to regional variability in the Indo-Pacific.

The results of this thesis are presented in five papers. In the first paper a new marine-based multi-proxy reconstruction for AMV is produced and presented, showing persistent multi-decadal variability 90 years further back in time than the instrumental records. The second paper evaluates multi-decadal variability in several proxy reconstructions of the Indian summer monsoon (ISM) and compares it with different AMV reconstructions, including the marine-based record from the first paper. Multi-decadal variability is found in ISM reconstructions back to the 15th century, but the relation with AMV is not clear. The AMV-ISM relation is further investigated in CMIP5 simulations in the third paper. While none of the models capture the observed significant AMV-ISM relation in the pre-industrial control simulations, one model simulates the observed correlation in the 20th century historical ensemble, indicating that the observed relation could be externally forced. In the fourth paper of this thesis it is found that changes in North Atlantic SSTs due to variations in the thermohaline circulation can impact variability in the equatorial Atlantic and change the inter-basin relation between Atlantic and Pacific Niños. The fifth paper introduces a novel approach for investigating large-scale impacts in a coupled model, by separating radiative forced and dynamically driven variability. Ensembles of partially coupled 20th century historical simulations indicate that AMV may not have been a key contributor in Northern Hemisphere surface trends on decadal timescales.

Collectively these papers indicate that the AMV is a persistent signal of the climate system, but the impacts on multi-decadal variability in the Indo-Pacific and the Northern Hemisphere may not be as strong as previous studies suggest. AMV can modulate interannual variability and inter-basin teleconnections in the tropics, but the correlation with the ISM could be due to external forcing, and the Pacific seems to make a larger contribution to decadal trends in Northern Hemisphere climate than the Atlantic.
List of papers


II. Sankar, S., L. Svendsen, G. Bindu, P. V. Joseph, and O. M. Johannessen, Teleconnections between Indian summer monsoon rainfall and Atlantic multidecadal variability over the last 500 years (*manuscript in preparation*).

III. Luo F., Y. Gao, L. Svendsen, N. Keenlyside, S. Li and T. Furevik, External forcing synchronizes Atlantic multidecadal variability and the Indian summer monsoon (*manuscript in preparation*).


V. Svendsen, L., N. Keenlyside, I. Bethke, and Y. Gao, Investigating the role of the Atlantic and Pacific in the early 20\textsuperscript{th} century warming trend (*manuscript in preparation*).

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1. Background and Motivation

The main objective of this thesis is to investigate the role multi-decadal variability in the Atlantic has for variability in the Indo-Pacific region and over the Northern Hemisphere. In the following I will present earlier studies that have motivated and form the scientific basis for the five papers that constitute this thesis. I will start with introducing Atlantic multi-decadal variability (AMV) and the large-scale impacts that have been recognized in earlier studies. Then I will focus on the specific impacts of AMV that this thesis deals with: decadal variability of the Indian summer monsoon (ISM), decadal modulations of interannual variability in the tropics, and finally the role of AMV in Northern Hemisphere surface temperature variability on decadal timescales.

1.1 Atlantic Multi-Decadal Variability

The observed multi-decadal variability of North Atlantic sea surface temperature (SST) anomalies is referred to as AMV or the Atlantic Multi-decadal Oscillation (AMO; e.g., Enfield et al. 2001). The AMV is characterized by basin-wide North Atlantic SST anomalies that vary between cold and warm periods of 3-4 decades each. The SST anomalies associated with AMV typically have a horseshoe pattern with the largest anomalies in the tropics and the eastern subtropical North Atlantic (Figure 1). An AMV-index (black line in Figure 2) is often defined as the area-averaged annual SST anomalies in the North Atlantic from the equator to 60°N (e.g., Enfield et al. 2001; Wyatt et al. 2012).

Both the drivers of AMV and how AMV impacts other regions are still unclear and heavily debated. When first detected, it was suggested that AMV was a signal of internal variability of the ocean; however later studies have suggested several externally forced components. While AMV can be related to internal variability of the thermohaline circulation in the Atlantic and the Atlantic meridional overturning circulation (AMOC) (e.g., Delworth et al. 1993; Delworth and Mann 2000; Latif et al.
some studies have suggested that AMV is a result of fluctuations in the atmospheric circulation, for instance by the North Atlantic Oscillation (NAO) (e.g., Eden and Jung 2001; Clement et al. 2015), or that external radiative forcing, natural, anthropogenic or a combination of the two, can modulate AMV (Otterå et al. 2010, Booth et al. 2012). But these results depend on the model and the definition of AMV (e.g., Zhang et al. 2013). Model studies are inconclusive and models vary widely both in terms of amplitude and frequency of AMV (Medhaug and Furevik 2011), while the spatial pattern is rather robust (Ting et al. 2011; Ba et al. 2014). This thesis is not a study of AMV, but rather of its impacts.

Figure 1: AMV pattern given by the correlation of low-frequency filtered SST anomalies with the AMV-index for the period 1871-2012, based on HadISST data (Rayner et al. 2003).

The short instrumental data only capture two cycles of AMV making it difficult to assess what drives AMV as well as AMV impacts. If AMV is internally driven it may be a persistent signal of the climate system and could have implications for climate prediction. Proxy reconstructions of North Atlantic climate have indicated some persistence of multi-decadal variability, although the records diverge prior to the instrumental era (e.g., Kilbourne et al. 2014). Most of these reconstructions are mainly based on land-records such as tree rings and ice cores (e.g., Gray et al. 2004; Mann et al. 2009; Knudsen et al. 2011). Land-based records assume a stable relation between SST and atmospheric temperatures, but this relation is not fully understood
(e.g., D’Arrigo et al. 2008). Proxy reconstructions are important for understanding possible drivers of AMV, but it is also essential for understanding long-term teleconnections by comparing them with reconstructions from other regions. The objective of Paper I in this thesis is to investigate if there is a persistent AMV-like signal in a reconstruction using only marine proxies, a more direct measure of SST compared to land-based records.

1.2 Impacts of Multi-Decadal Variability in the Atlantic

Previous studies have suggested that AMV is important for climate variability globally (e.g., Schlesinger and Ramankutty, 1994; Delworth and Knudsen 2000; Zhang et al. 2007; Steinman et al., 2015). AMV has been linked to European and North American summer climate (e.g., Enfield et al. 2001; McCabe et al. 2004; Sutton and Hodson 2005; Sutton and Hodson 2007; Wyatt et al. 2012), Arctic sea ice cover (e.g., Kinnard et al. 2011; Miles et al. 2014), and Atlantic hurricane activity (e.g., Goldenberg et al. 2001; Knight et al. 2006; Zhang and Delworth 2006; Wang et al. 2012).

AMV has also been related to summer rainfall in African Sahel from June to August through a meridional shift of the Inter-tropical convergence zone (ITCZ); during a positive AMV the ITCZ is shifted northward enhancing precipitation over Sahel (e.g., Folland et al. 1986; Knight et al. 2006; Zhang and Delworth 2006; Mohino et al. 2010; Ting et al., 2011; Wang et al. 2012). Similarly a cold AMV has been found to enhance rainfall in North Eastern Brazil from March to May also through a shift of the ITZC (e.g., Knight 2006; Ting et al. 2011). Multi-decadal variability in the South Asian summer monsoon could also be impacted by AMV, with a positive correlation observed between SST anomalies in the North Atlantic and boreal summer rainfall over South Asia (e.g., Goswami et al. 2006; Zhang and Delworth 2006; Li et al., 2008; Ting et al. 2011). Recent studies have also suggested that multi-decadal variability in the Atlantic can modulate interannual variability and inter-basin
coupling in the tropics (e.g., Chen et al. 2010; Polo et al. 2013; Kang et al. 2014; Martin-Rey et al. 2014).

So-called freshwater hosing experiments with global climate models where a large amount of freshwater is artificially added to the North Atlantic resulting in weaker AMOC and a cooler North Atlantic, have also shown that basin-wide changes in the North Atlantic can have far-reaching effects impacting for instance the mean state of the tropical Pacific, but also interannual variability such as the El Niño-Southern Oscillation (ENSO) (e.g., Timmermann et al. 2005; Zhang and Delworth 2005; Dong and Sutton 2007, Rashid et al. 2010), the Atlantic Niño (e.g., Polo et al. 2013), ISM rainfall (e.g., Lu and Dong 2008) and the relation between ENSO and the ISM (e.g., Chen et al. 2010). These studies relate to AMV impacts to the extent that a weaker AMOC results in similar SST changes as those associated with the observed AMV.

This thesis deals with the impacts of AMV especially focusing on the Indo-Pacific region. Figure 2 illustrates the three impacts that are the focus on this thesis: the relation to the ISM rainfall (yellow line), the inter-basin connection between the

Figure 2: Observed normalized and de-trended low-frequency filtered Northern Hemisphere surface temperature (blue line) from the GISTEMP data (Hansen et al. 2010), All-India monsoon rainfall index (yellow line) from IITM (Parthasarathy et al. 1994), the AMV-index (black line), and the 21-year running correlation between Atl3-index (3°N-3°S, 20°W-0°W) and Nino3-index (5°N-5°S, 90°W-150°W) (green line) from the HadISST data (Rayner et al. 2003).
tropical Atlantic and Pacific (green line), and Northern Hemisphere surface temperatures (blue line). In the following I will present recent studies that motivated this work.

1.3 AMV and the Indian Summer Monsoon

A significant correlation between AMV and ISM rainfall has been observed (e.g., Goswami et al. 2006), with periods of warm (cool) North Atlantic SSTs associated with periods of more (less) ISM rainfall (yellow line in Figure 2). Whether the Atlantic is driving multi-decadal variability in the ISM is uncertain and a mechanism for the possible connection is disputed.

The ISM is the rainy season over India. During the months from June to September India receives almost 80% of its annual rainfall (Figure 3; Tyagi et al. 2012). Based on the seasonal reversal of the wind direction between boreal winter and summer due to annual variations in the incoming solar radiation and the different heating capacity over land and ocean, the ISM can be seen as an enormous “sea breeze”. It can also be viewed as a seasonal migration of the ITCZ (e.g., Gadgil 2003; Privé and Plumb, 2007; Joseph 2014). During spring and summer South Asia warms more than the Indian Ocean, and a meridional pressure gradient in the upper troposphere between the Indian continent and the Southern Indian Ocean is established and air flows southwards in the upper troposphere. Due to mass continuity warm moist air from over the Indian Ocean flows northwards in over South Asia and the Indian Peninsula. The Coriolis force turns the cross-equatorial low-level winds over the Arabian Sea crossing the Indian peninsula from west to east (Figure 3; Tyagi et al. 2012). Simultaneously the monsoon trough over central India brings low-pressure systems inn from the Bay of Bengal (Gadgil 2003). A dryer ISM has typically a weaker monsoon circulation with a weaker Somali Jet (e.g., Findlater 1969; Joseph 2014), weaker Hadley circulation and Walker circulation (e.g., Webster et al. 1998), and cooler temperatures over the Tibetan Plateau reducing the meridional temperature
gradient between the Indian Ocean and Eurasia (e.g., Goswami and Xavier 2005). These anomalies are typically reversed for a wetter ISM.

Figure 3: Seasonal cycle of monthly mean rainfall in mm over India (left panel) derived from the Climatic Research Unit (CRU) data set (provided from the NCAS British Atmospheric Data Centre; retrieved from https://climatedataguide.ucar.edu/climate-data/cru-ts321-gridded-precipitation-and-other-meteorological-variables-1901; Harris et al. 2014), and the mean surface wind pattern (right panel) during the ISM (JJAS) in m/s from NCEP/NCAR reanalysis data (provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their web site http://www.esrl.noaa.gov/psd), for the period 1981-2010 (Kalnay et al. 1996).

1.3.1 The AMV-ISM Link in Proxy Reconstructions

Since the instrumental records are short compared to the multi-decadal timescales of AMV, proxy reconstructions are used to assess the persistence of the AMV-ISM relation. Paleo-climate records provide evidence that low-frequency variation in SST in the Atlantic concurred with variations in the ISM, for instance there is evidence for changes in the Asian monsoon that correspond in timing to Dansgaard/Oeschger events (Burns et al., 2003) and smaller interglacial changes in the North Atlantic (Gupta et al. 2003, Feng and Hu 2008). There is also evidence of multi-decadal variability in annual resolution proxy record of regional rainfall in India (e.g., Berkelhammer et al. 2010; Borgaonkar et al. 2010).

Rainfall approximations from one region in India are not necessarily representative for the monsoon as a whole (Normand 1953). Individual proxy reconstructions include components of local variability, and therefore multi-proxy reconstructions
using proxies from several different sites are preferred. For the ISM there are several proxy reconstructions, but few of them are publically available. There are a few records from India, and also a couple of marine records from both the Indian and the Pacific Ocean related to ISM rainfall (e.g., Borgaonkar et. al. 2010; Yadava et al., 2004; Sinha et al., 2007; Chakraborty et al., 2012). In addition to individual proxies, a 700-year drought atlas for the whole Asian monsoon region, the Monsoon Asia Drought Atlas (MADA), is available (Cook et al., 2010). The MADA is a seasonally resolved drought reconstruction including 534 grid points based on 327 tree ring series from Asia. However only four tree ring records are included from sites in India, and only two grid points are significantly correlated with the observed ISM rainfall (not shown).

Low-resolution proxy records indicate that the variability in the North Atlantic and the ISM could be related on longer timescales, and individual records do indicate persistence in multi-decadal variability in the ISM. A multi-proxy analysis of multi-decadal variability of the ISM is missing in the literature, in addition to a comparison with AMV reconstructions. The ISM reconstructions that are available are evaluated and compared with multi-proxy reconstructions of AMV in Paper II.

1.3.2 Dynamical Connection between AMV and the ISM

It is still unclear if the observed correlation between AMV and ISM indicates causality and how the Atlantic could then potentially influence the ISM. Goswami et al. (2006) demonstrated that the quasi 60-year oscillation of the ISM varies coherently with AMV in observations, as well as the low-pass filtered ENSO but with opposite signs. When the AMV was positive there were positive tropospheric temperature anomalies over Eurasia increasing the meridional tropospheric temperature gradient (MTG) between the Indian Ocean and Eurasia. This could delay the withdrawal of and increase the intensity of the ISM. Goswami et al. (2006) found that the AMV influenced the tropospheric temperature through the NAO acting as an atmospheric bridge. A positive summer NAO-index was associated with a similar pattern as a positive AMV, and the changes of the winds and storm tracks associated
with NAO variability could be responsible for changes in the ISM (Goswami et al. 2006). Feng and Hu (2008) also found that temperatures over the Tibetan Plateau warm and the Indian Ocean cools during positive AMV phases, consistent with the strengthening of the MTG. A stronger MTG could shift the ITCZ northwards and enhance low-level winds, bringing more moisture over peninsular India, increasing ISM rainfall (Feng and Hu 2008). Similarly Wang et al. (2009) documented that during a positive AMV, northern India had warmer surface temperatures, and southern India had cooler temperatures. The boundary between these two temperature lobes moves northwards through the year, and there was a negative correlation between the AMV index and South Asian surface temperatures, as well as an increase rainfall in boreal summer (Wang et al. 2009).

Model studies both reveal possible mechanisms for an AMV-ISM connection and can indicate whether the connection is a part of natural internal climate variability or externally forced. Decadal trends in the ISM rainfall have been related to anthropogenic aerosol emissions (Bollasina et al. 2011) or solar variability (Meehl et al. 2003; Bhattacharyya and Narasimha 2005). However results from fully coupled model simulations with constant external forcing suggest that the observed correlation between AMV and the ISM can also be due to internal variability of the climate system (e.g., Msadek and Frankignoul 2009; Luo et al. 2011).

The skill in simulating decadal variability of the ISM is found to be better in atmospheric general circulations models (AGCMs) than in coupled GCMs, while interannual variability is better simulated in coupled GCMs, suggesting that ocean-atmosphere coupling is important for the interannual variability, while decadal variability is mainly an atmospheric signal or that the SST pattern is important (Kucharski et al. 2009). However studies using both atmospheric and coupled GCMs have in general found that SST anomalies in the North Atlantic associated with a positive AMV lead to an intensification of the lower level Somali Jet and the south westerly surface winds from the Indian Ocean across India due to a northward shift of the ITCZ and a strengthening at the MTG due to tropospheric heating over Eurasia.
(Zhang and Delworth 2006; Li et al. 2008; Lu et al. 2006; Wang et al. 2009; Luo et al. 2011). Similar responses in the Asian summer monsoon circulation have been found for a weaker AMOC and associated negative SST anomalies in the North Atlantic (Lu and Dong 2006; Msadek and Frankignol 2009). However models are not consistent in simulating a response in the ISM rainfall, and several studies only find a significant rainfall response in boreal autumn during the ISM withdrawal phase (Lu et al. 2006; Wang et al. 2009; Luo et al. 2011). Conversely instrumental data show that the wet and dry monsoon decades are mostly related to the amount of July rainfall, the peak monsoon month, rather than the monsoon withdrawal phase (Tyagi et al. 2012).

AMV might impact the ISM through a Rossby wave train from the North Atlantic across Eurasia. North Atlantic SST anomalies can result in baroclinic instability over western Europe, and this perturbation could induce a Rossby wave across Asia. This can lead to a change in the MTG over India and enhance the South Asian high, increasing upper-level divergence and low-level convergence, in turn changing the intensity of the ISM (Li et al. 2008; Luo et al. 2011). A meridional shift of the ITCZ in the Atlantic associated with AMV can also excite a Gill-type response in the eastern Pacific changing the strength of the Walker circulation and consequently the ISM (Lu and Dong 2008; Zhang and Delworth 2005).

However studies disagree on the sign of the response in the ISM. Both Sutton and Hodson (2007) and Li et al. (2008) investigated the role of SST anomalies in the tropical and extra-tropical part of the AMV signal separately in driving ISM variability using different AGCMs with prescribed SST anomalies. Both studies found that tropical North Atlantic SST anomalies had a negative correlation with the ISM, while extra-tropical North Atlantic SST anomalies had a positive correlation with the ISM. Warm anomalies in the tropical North Atlantic inhibit convection in regions outside the tropical Atlantic and weaken the Somali Jet, suppressing rainfall over India and the Bay of Bengal (Sutton and Hodson 2007). Sutton and Hodson (2007) found the greatest response in the ISM from tropical Atlantic SST anomalies,
while Li et al. (2008) simulate the greatest response from extra-tropical Atlantic SST anomalies. This resulted in an overall opposite relation between AMV and the ISM in these two studies, with the results in Sutton and Hudson (2007) also opposite of the observed AMV-ISM relation. Air-sea coupling in the tropics might therefore be important to reverse the sign of the response, especially in the Indian Ocean (Sutton and Hodson 2007; Kucharski et al. 2007).

These results emphasize the model dependency of such studies. There is also little agreement in the correlation between AMV and ISM rainfall between the models that participated in the Coupled Model Intercomparison Project Phase 3 (CMIP3), and the correlation is weak in the ensemble mean for the 20th century historical simulations (Ting et al. 2011). GCMs still have difficulties simulating the ISM. These deficiencies have been related to low resolution and the representation of orography, convective parameterizations, and biases in the mean state of the tropics (Turner et al. 2011), but ISM rainfall patterns seem to be better simulated in coupled GCMs with stronger correlations between the ISM rainfall and equatorial Pacific SSTs (Sperber et al. 2013). The weak connection between AMV and the ISM in CMIP3 may be attributed to the response in the Pacific to an AMV signal that is opposite compared to observations: in observations warm North Atlantic SSTs are associated with small negative SST anomalies in the equatorial Pacific, while in the CMIP3 ensemble the equatorial Pacific SST anomalies are positive and significant (Ting et al. 2011). The Pacific response to AMV may be important for simulating the connection with the ISM, and the ability to simulate tropical atmosphere-ocean coupling could be crucial for these teleconnections.

Goswami et al. (2006) found that decadal variability in ISM rainfall is modulated by two teleconnections patterns; first a tropical teleconnection by the low-pass filtered ENSO involving shifts of the Walker circulation influencing the regional monsoon Hadley circulation, second an extra-tropical teleconnection where AMV influenced the monsoon through changing the tropospheric temperatures over Eurasia. The regression pattern in Figure 4 shows that observed multi-decadal variability of the
ISM rainfall is significantly correlated with multi-decadal variability in both the tropical Pacific and the North Atlantic. The CMIP5 models do not seem to be able to simulate both these correlations (Figure 5). Paper III is a further investigation of the AMV-ISM relation in the latest version of the CMIP models (CMIP5; Taylor et al. 2012), the drivers for this relation, and whether the relation is internal variability or externally forced.

Figure 4: Regression of low-pass filtered global SSTs (HadISST) onto the ISM rainfall index (from IITM) for the period 1871-2004. Black dots indicate significant values at a 5 % level for the effective degrees of freedom.

Figure 5: Scatter plot of the correlation between ISM rainfall and AMV (1.axis) and ISM rainfall and Inter-decadal Pacific Oscillation (IPO) given by SST anomalies in the equatorial Pacific (2.axis) for 20 CMIP5 models (colored dots), and instrumental data (black x). Values beyond the dashed lines are significant at a 5 % level.
AMV could also potentially modulate interannual variability of the ISM. During decades of less ISM rainfall interannual variability of the ISM and the frequency of dry years increased (Joseph et al. 2013). Historically dry and wet ISM years have been related to the Southern Oscillation and ENSO events (e.g., Walker 1924; Rasmussen and Carpenter 1983; Ju and Slingo 1995), but this relation is not linear (Webster et al. 1998, Torrence and Webster 1999), and Kumar et al. (1999) found that this relation has weakened in recent decades. The tropical Atlantic can also influence the ISM through similar mechanisms as ENSO (Kucharski et al. 2007; 2008). Decadal changes in the Atlantic could therefore potentially impact the ISM through changing tropical Atlantic variability (e.g., Polo et al. 2013), ENSO variability (e.g., Timmermann et al. 2005; Zhang and Delworth 2005; Dong and Sutton 2007) or the relation between ENSO and the ISM (Chen et al. 2010).

1.4 Interannual Variability Potentially Modulated by AMV

Enfield et al. (2001) suggested that AMV affects rainfall variability over North America associated with ENSO, modulating interannual variability in this region. Recent studies have also suggested that multi-decadal variability in the Atlantic can modulate interannual variability and inter-basin coupling in the tropics (e.g., Kang et al. 2014; Martín-Rey et al. 2014). There is also observational evidence that the relation between tropical Atlantic variability associated with an Atlantic Niño and ENSO has strengthened since the 1970s in phase with a negative AMV (e.g., Rodriguez-Fonseca et al. 2009; Martín-Rey et al. 2014), also apparent in Figure 2 (green line). These modulations could in turn impact for instance the ISM, through its relation to the tropical Atlantic (Kucharski et al. 2007; 2008) or ENSO (Walker 1924; Rasmussen and Carpenter 1983; Ju and Slingo 1995)

ENSO is a coupled atmosphere-ocean oscillation in the tropical Pacific (Bjerknes 1969, Philander 1990), and presents itself as SST anomalies in the equatorial Pacific accompanied by a change in the zonal pressure gradient across the Pacific. As the dominant mode of natural climate variability on interannual timescales, it can affect
climate globally as well as being important for the ISM noted above. Similar oscillatory interannual variability also exists in the Atlantic, referred to as the Atlantic zonal mode or Atlantic Niño (e.g., Zebiak 1993). The variability here is less than in the Pacific and has a different annual cycle. While ENSO peaks in boreal winter, the Atlantic Niño peaks in summer (Figure 6).

Figure 6: Standard deviation of SST (HadISST) in the tropical Pacific during boreal winter (left panel) and in the tropical Atlantic during boreal summer (right panel).

As mentioned earlier, several waterhosing experiments or similar studies prescribing SSTs in the Atlantic have found that basin-wide changes in North Atlantic can change the mean state of the tropical Pacific and ENSO variability (Dong and Sutton 2002; Timmermann et al. 2005; Zhang and Delworth 2005; Dong and Sutton 2007; Timmermann et al. 2007). A weakening AMOC induces a temperature dipole in the tropical Atlantic shifting the ITCZ southwards, and the atmospheric response changes the mean state of the tropical Pacific (Dong and Sutton 2002; Zhang and Delworth 2005; Dong and Sutton 2007; Timmermann et al. 2007). However not all models agree on if this will increase or decrease ENSO variability (Timmermann et al. 2007). SSTs in the tropical North Atlantic are also found to trigger ENSO events (Ham et al. 2013). Similarly a weakening AMOC could change the variability in the Atlantic Niño through the impact on ENSO (e.g., Polo et al. 2013).

Tropical Atlantic variability can also impact the tropical Pacific on interannual timescales, with SST variations in the tropical Atlantic impacting ENSO through atmospheric teleconnections (Dommenget et al. 2006; Jansen et al. 2009; Rodriguez-Fonseca et al. 2009; Frauen and Dommenget 2012). The correlation between SST
anomalies in the central equatorial Pacific and the central equatorial Atlantic is strongest when the Atlantic is leading the Pacific by 6 months (Keenlyside and Latif 2007; Rodriguez-Fonseca et al. 2009). Warm (cold) SST anomalies in boreal summer associated with an Atlantic Niño (Niña) event can drive a shift in the Walker circulation, strengthening (weakening) the equatorial trade winds over the Pacific leading to La Niña-like (El Niño-like) SST anomalies in the Pacific through the positive Bjerknes feedback (Bjerknes 1969). These studies suggest that the variability in the Atlantic Ocean may be important for both increasing the understanding of ENSO variability and predicting ENSO (Rodriguez-Fonseca et al. 2009; Ding et al. 2012; Keenlyside et al. 2013).

Previous studies have indicated a strengthening of the relationship between Atlantic Niño events and ENSO since the 1970s (e.g., Keenlyside and Latif 2007; Rodriguez-Fonseca et al. 2009). This could be due to stochastic changes (Ding et al. 2012), or changes in the background state of the Pacific and the Atlantic (Rodriguez-Fonseca et al. 2009). Martín-Rey et al. (2014) suggests that the AMV is the modulator for the variations of the inter-basin coupling between Atlantic Niños and ENSO events, with a negative AMV strengthening the coupling. Paper IV investigates how this tropical Atlantic-Pacific coupling is influenced by multi-decadal changes in the Atlantic in a waterhosing experiment.

1.5 Global Surface Temperature Variability and the AMV

As seen in the previous sections multi-decadal variability in the North Atlantic can impact climate variability globally. Both global and Arctic temperatures in the recent 150 years have exhibited multi-decadal variability with an early 20th century warming from 1910-1940 with a following cooling period before a second pronounced warming period from the late 1970s (the blue line in Figure 2 illustrates this variability in the Northern Hemisphere). Global and Arctic multi-decadal temperature variability has been related to variability in the Atlantic (e.g., Schlesinger and Ramankutty 1994; Delworth and Mann 2000; Zhang et al. 2007; Steinman et al.
2015), however the Pacific has also been suggested as a driver (e.g., Keenlyside and Ba 2010; Koseka and Xie 2013; Meehl et al. 2013; Dai et al. 2015), as well as anthropogenic or natural external forcing (e.g., Tett et al. 1999; Stott et al. 2000; Tett et al. 2002; Broccoli et al. 2003; Hegerl et al. 2003, Meehl et al. 2004).

For instance volcanic aerosols could lead to a cooling, and the early 20th century warming period has been in part attributed to low volcanic activity. In addition there is evidence that there was an increase of solar insolation during this period. Atmospheric greenhouse gas concentrations have increased, intensifying since the 1970s consistent with the later warming period. All these external factors could have played a role in global surface temperature variability (Tett et al. 1999; Stott et al. 2000; Broccoli et al. 2003; Hegerl et al. 2003, Meehl et al. 2004). Yet global coupled climate models are often not able to simulate the observed phasing of the temperature variability even when the models include external forcing, suggesting that low-frequency variability in the ocean, for instance AMV or the Pacific Decadal Oscillation (PDO; e.g., Mantua and Hare 2002), could be important.

A related topic is the slowdown of the global warming trend in the last decade since the end of the 1990s, and also here there is an ongoing debate whether this is externally forced or due to internal variability. Several studies have suggested a cooling of the Pacific as the cause for this Hiatus (Salomon et al. 2010; Koseka and Xie 2013; Meehl et al. 2013; Trenberth and Fusallo 2013; Fyfe and Gillett 2014; England et al. 2014). Conversely Chen and Tung (2014) suggested the importance of the Atlantic with the heat content of the deeper layers of the Atlantic increasing in the recent decade. However what drives low-frequency variability in the Atlantic and Pacific Ocean is not clear, and some studies have for instance concluded that external forcing can modulate the AMV (Otterå et al. 2010; Booth et al. 2012), indicating the possibility that external forcing could be driving both AMV and global surface temperatures. The role of AMV in multi-decadal global surface temperature variability is the topic of Paper V.
2. Open Questions and Main Objectives

The main objective of this thesis is to investigate impacts of multi-decadal variability in the Atlantic on global variability, and especially on variability in the Indo-Pacific region. The previous chapter showed that multi-decadal variability in the North Atlantic could impact global temperatures and regional climate variability, such as in the tropical Pacific and Atlantic and the ISM, but several open questions remain on this topic.

Answering these questions will increase our knowledge of large-scale climate interaction that can be important for understanding regional climate variability, as well as indentify weaknesses and possible areas of improvement in coupled GCMs, improving the potential for skillful decadal climate predictions and reducing uncertainties in regional climate change projections. Analysis of both proxy reconstructions and GCMs can help us address these questions by recognizing possible links in the climate system and decompose internal variability and externally forced changes.

This thesis focuses on the following underlying research questions:

1. How did North Atlantic SSTs vary on multi-decadal timescales prior to the instrumental records?
2. How is AMV connected to the ISM on multi-decadal timescales?
3. How do multi-decadal changes in the North Atlantic impact inter-basin couplings in the tropics?
4. What part does the North Atlantic play in multi-decadal variability of Northern Hemisphere surface temperatures?
3. Summary of Results

All papers in this thesis relate to the main topic of multi-decadal variability in the North Atlantic, with focus on the impacts on decadal timescales. Paper I deals with the first research question, reconstructing a new multi-proxy marine-based AMV-index to investigate the persistence before the instrumental period. Paper II and III are concerned with research question 2, investigating the relation between AMV and the ISM, in proxy records in paper II and in CMIP5 models in paper III. Paper IV contributes to the topic of research question 3 on how the decadal changes in the AMOC and associated mean state changes in the North Atlantic can modulate inter-basin teleconnections in the tropics. Paper V considers research question 4 suggesting that perhaps the AMV is not the main driver in multi-decadal variability in Northern Hemisphere temperatures, especially for the early 20th century warming. The five papers that constitute this thesis are summarized below, highlighting the key findings.


The motivation for Paper I was that there are several AMV reconstructions that are heavily used in the scientific literature, both when comparing with other reconstructions and to validate model simulations. These few reconstructions are mostly composed of land-based records, which depend on a stable relation between North Atlantic SST and atmospheric temperatures over surrounding landmasses, which might not be the case. In Paper I a new AMV-reconstruction is made using only marine-based proxies. We find that by combining several coral records with annual resolution from the tropical Atlantic using principle component analysis, we are able to capture the observed multi-decadal variability. This variability also persists throughout the multi-proxy reconstruction back to year 1781 suggesting that AMV is persistent, but differs slightly in the timing from other widely used AMV-
reconstructions. The results motivate the use of more marine-based proxies for such reconstructions.

Key findings:

• Multi-decadal variability is found as a common component of variability in coral records from the tropical Atlantic
• AMV is persistent at least back to year 1781.

Paper II: Teleconnections between Indian summer monsoon rainfall and Atlantic multidecadal variability over the last 500 years, Sankar, S., L. Svendsen, G. Bindu, P. V. Joseph, and O. M. Johannessen (manuscript in preparation).

Although AMV and the ISM have been linked together in observations, it is not clear if this link is persistent. Paper II evaluates and compares decadal variability in several ISM proxy reconstructions with annual resolution. Although published, some of these records are not publically available and therefore have never been compared before. In addition these ISM reconstructions are for the first time compared with several AMV reconstructions, including the marine-based AMV reconstruction presented in Paper I. Although the analysis shows persistent variations of dry and wet ISM decades before the instrumental record begins, the correlation with AMV, that is significant in the instrumental period, is not clear. The available data for both the Atlantic and the ISM are scarce and have discrepancies, and one of the main conclusions of this study is that more high quality data needs to be available, as well as highlighting the importance of using several proxy records for such comparisons. However, the correlation in the observed records has also weakened in the recent decades suggesting that the observed AMV-ISM link might not be stable.

Key findings:

• Multi-decadal variability in the ISM related to the frequency of drought years is persistent in proxy reconstructions back to the 15th century
• The observed correlation between AMV and ISM rainfall is not stable weakening in the recent decades, and possibly before the 19th century as well.

Paper III investigates if the CMIP5 models are able to simulate the observed relation between AMV and multi-decadal variability of the ISM, and if this link is internal climate variability or externally forced. After an evaluation of both AMV and ISM in 25 CMIP5 models and selecting the top five of these for further analysis, it is found that only one of these models (GFDL-CM3) simulates the observed AMV-ISM correlation in the historical 20th century simulations as observed. None of the models we evaluated had the observed correlation in the pre-industrial control simulation, indicating that if the connection is internal climate variability, the models are not able to simulate this. An evaluation of the GFDL-CM3 model showed that the observed relation could be externally forced by modulating both AMV and the ISM through a response in upper tropospheric temperatures in the subtropics.

Key findings:
• The observed AMV-ISM link is not a robust feature in CMIP5 models
• The observed correlation between AMV and the ISM could be due to a response in both the ISM and North Atlantic SSTs to external forcing


Several so-called waterhosing experiments where freshwater is artificially added to the North Atlantic have shown a reduction in the AMOC cooling the North Atlantic, as well as significant changes in the mean state of the Pacific and ENSO variability. Paper IV investigates a similar waterhosing experiment with the Bergen Climate Model, and shows that a more moderate weakening of the AMOC does not significantly change the mean state of the tropical Pacific, but increases equatorial Atlantic SST variability and strengthens the connection between the equatorial
Atlantic and Pacific, increasing the ENSO frequency. The strengthening of the Atlantic–Pacific relation when the Atlantic is colder than normal is consistent with observed records where this relation has varied in phase with the AMV. However the reason for the shift in the ENSO frequency is still uncertain. Although not mentioned specifically in the paper, we also see a significant increase in the interannual connection between tropical Atlantic summer SSTs and Indian summer monsoon rainfall (see Figure 8 a and b in Paper IV) when the Atlantic is colder than normal, with warm anomalies in the equatorial Atlantic associated with less ISM rainfall.

Key findings:
• A weakening AMOC can increase variability in the tropical Atlantic
• AMOC variability can modulate the inter-basin connection between Atlantic Niño and Pacific ENSO-events


There is an ongoing discussion about the relative contribution of natural and anthropogenic external forcing and internal climate variability on decadal global temperature variability. Paper V investigates the portion of Northern Hemisphere surface temperature variability that is not directly driven by radiative forcing, to see if multi-decadal variability in the Atlantic is a key driver in this signal. Four six-member ensembles of partially coupled experiments were performed with the Norwegian Earth System Model (NorESM), with prescribed momentum flux anomalies from reanalysis data to the global ocean, to the Atlantic or to the Indo-Pacific. By prescribing momentum flux, SST variability is constrained to the observed in the respective regions, while the model is still thermodynamically coupled. Since all the simulations include transient 20th century historical forcing, these experiments represent a new approach for separating radiative forced and dynamically driven variations. We find that external forcing accounts for about half of the early 20th century warming in the Northern Hemisphere and for the Arctic
specifically, and dynamically driven variability accounts for the other half. Decadal variability in the Pacific is well simulated, while the amplitude of the simulated AMV is too small and the periodicity is shorter than observed. Even though multi-decadal variability in the Atlantic is not well simulated we still manage to simulate the early 20\textsuperscript{th} century warming in the Northern Hemisphere due to the phasing of decadal variability in the Pacific. We conclude that the phasing of AMV may not be a key contributor to the early 20\textsuperscript{th} century warming.

Key findings:

\begin{itemize}
\item PDO can impact Northern Hemisphere and Arctic surface temperature variability on decadal timescales, especially contributing to the early 20\textsuperscript{th} century warming
\item AMV is not central to the early 20\textsuperscript{th} century warming in the Northern Hemisphere.
\end{itemize}

3.1 Main Conclusion

The results from the five papers in this thesis lead to the following conclusions. Multi-decadal variability in the Atlantic seems to persist before the short instrumental record. This variability can impact the Indo-Pacific region, by for instance strengthening the ISM or the inter-basin Atlantic-Pacific Niño relation. However while there seems to be a connection between the ISM and AMV in observations, this connection is neither clear in the proxy records before the 19\textsuperscript{th} century, nor reproducible in state-of-the-art coupled GCMs. However one model can capture the observed correlation in the 20\textsuperscript{th} century historical simulations as a response to external forcing, indicating that other models might be less sensitive to the prescribed external forcing. On a larger scale the AMV also seems less important for driving the early 20\textsuperscript{th} century warming in the Northern Hemisphere and Arctic surface temperatures; our results suggest a dominating role of the Pacific. In short, the persistent AMV can potentially modulate interannual to multi-decadal variability and inter-basin teleconnections in the tropics, but the AMV plays a minor role in both decadal trends in Northern Hemisphere temperatures and the ISM, with the multi-decadal variability in the ISM possibly driven by external forcing.
4. Discussion and Future Perspectives

This thesis synthesizes studies on how AMV impacts the Indo-Pacific and Northern Hemisphere on interannual to multi-decadal timescales. The approach here is based on the hypothesis that multi-decadal variability in the Atlantic can be a key driver of such variability. However there are still great uncertainties and limitations related to this topic, due to for instance biases in coupled GCMs and their range in climate sensitivity and level of internal variability, as well as limited data coverage on these timescales. This thesis has aimed to deal with some of these limitations, by investigating simulations of these interactions in state-of-the-art coupled GCMs and extending the data coverage using proxy reconstructions. The results of this thesis do not fully answer the questions proposed in Chapter 2, but contribute to the knowledge on these topics. This thesis also motivates further research and some of these ideas are presented in the following.

The new marine-based AMV reconstruction presented in Paper I suggests that AMV has persisted prior to the instrumental record. However this new reconstruction is still only 90 years longer than the instrumental records, and a longer multi-proxy reconstruction is preferred. Longer records from the Atlantic are available, but these have lower resolution. These lower resolution records, preferably covering a larger area of the Atlantic, could be combined using a similar method as in Paper I to assess multi-decadal variability even further back in time.

A similar method of combining marine records can be used to assess variability in other regions as well. These multi-proxy reconstructions from different regions can then be compared to determine common variability and to investigate if observed teleconnections hold prior to instrumental records. The method could also be used for land-based records over India to estimate variability in the ISM rainfall. Historical ship records of wind direction and speed from the Atlantic have been used to reconstruct the West African monsoon back to year 1790 (Gallego et al. 2015), and similarly ship records from the Indian Ocean could be used for ISM circulation
reconstructions. Paper II highlights the fact that very few records from the ISM region are available, and Paper II is therefore only a comparison or these records. More records from India and the Indian Ocean are needed for the method used in Paper I to be useful to reconstruct large-scale features of the South Asian monsoon.

Instrumental records and several model studies have linked AMV to the ISM. Multi-decadal variability in the North Atlantic seems persistent, but it is not clear if the observed link to the ISM is also persistent. The comparison of ISM and AMV proxy reconstructions presented in Paper II showed that even though records from both regions exhibit multi-decadal variability, they may not always be in phase. However, based on the records that are presently available the persistence of the observed AMV-ISM link is still unclear, and deserves further investigation as well as better quality data.

Furthermore most of the state-of-the-art coupled models are not able to simulate the observed relation between AMV and the ISM, and Paper III found that only one model could simulate this in the 20th century all-forcing ensemble. The analysis of the CMIP5 models in Paper III presents only a first step in the analysis. Further analysis has to be done to understand why most CMIP5 models cannot simulate this relation. Preliminary results on this topic show that the simulation of variability in the Pacific or aerosol indirect effects might be important. A new CMIP will also be available in the near future and it will be interesting to see if this new suite of models performs differently.

The results from Paper III indicate that external forcing can modulate multi-decadal variability in both Atlantic SSTs and the ISM rainfall, in phase with global surface temperatures, giving us the observed correlation between the two regions. Several earlier studies have found that external forcing is important for modulating multi-decadal variability in global surface temperatures and AMV, as well as driving decadal trends in the ISM. For instance aerosol forcing specifically have been related to both AMV (Booth et al. 2011) and trends in the ISM since the 1950s (Bollasina et
External forcing is also found to be one of the main drivers in Northern Hemisphere surface temperatures on decadal timescales in Paper V, but dynamically driven variability in the Pacific seems equally important. Multi-decadal variability in both the ISM and AMV in the GFDL-CM3 model analyzed in Paper III are also related to SSTs in the North Pacific. However since GFDL-CM3 is the only model that captures the observed AMV-ISM relation of those analyzed in Paper III and since the simulated ISM is not significantly correlated with the observed even though it is correlated with AMV, this model might be too sensitive to external forcing. Whether multi-decadal variability in Atlantic SSTs and ISM rainfall is due to external forcing or internal climate variability is still unclear as results are model dependent.

Paper V introduced a novel approach to disentangle these factors by separating radiative forced and dynamically driven variability. Earlier studies have shown that both the AMV and variability in the Pacific can be important for driving decadal global temperature variations. Paper V concludes that by only driving the Pacific we can simulate the early 20th century warming in the Northern Hemisphere and especially in the Arctic, and the observed AMV is not crucial for these global variations. We used the 20CR to perform these experiments, and similar ensembles with other long reanalysis products (e.g., ERA-20C) could be performed to assess the uncertainties related to the prescribed wind stress. The ensembles could also be extended to present so that they can be used to investigate the recent hiatus period as well. In addition the same experimental setup can be used to force smaller regions. The setup has shown to be able to simulate the observed ENSO events, and by only prescribing momentum flux anomalies in the tropical Pacific, these experiments can be used to assess ENSO teleconnection patterns and impacts.

While the phasing of PDO is well simulated in the experiments used in Paper V, the AMV is not constrained to observations after the 1950s. Efforts should be made to improve the experimental design for simulating the AMV, for instance by prescribing heat flux anomalies in this region. These experiments could then also be used to assess AMV teleconnections to for instance the ISM, and impacts on the inter-basin
relation between the tropical Atlantic and Pacific. Recent studies have shown that while AMV and the ISM are significantly correlated in observations, there are regional differences with rainfall in some regions in India relating more to AMV, while other regions are perhaps more related to PDO or the Indian Ocean (e.g., Joshi and Rai 2014). This could also potentially be assessed using the same experimental design.

Several studies from the last decade have found that changes in the mean state of the Atlantic, such as those induced by a weaker AMOC or a negative AMV, can lead to a change in the mean state of the Pacific and a change in ENSO variability and amplitude. Polo et al. (2013) also found that the increased variability in the Pacific feeds back onto the tropical Atlantic increasing variability here. Paper IV, in contrast, shows that a weakening AMOC can increase the interannual variability in the tropical Atlantic, strengthen the inter-basin connection between the tropical Atlantic and Pacific, increasing the ENSO frequency, while the mean state of the Pacific exhibits no significant change. In addition to the more modest AMOC change compared to similar studies, the model used in Paper IV is flux-adjusted with a well-defined Atlantic cold tongue, which can strengthen the skill in tropical teleconnections on interannual timescales (Turner et al. 2005). However decadal variability in running correlations are innate statistical features of two random time series with interannual variability (Gershunov et al. 2001), and decadal changes in ENSO frequency can also be driven solely by internal variability (Wittenberg et al. 2009; 2014). The results in Paper IV are from one realization of one model, although to some extent consistent with observations (Martín-Rey et al. 2014), should be tested with similar experiments with other models, or in an ensemble with the same model (but technically difficult as the model is no longer in use).

Results in this thesis showed that AMV could be persistent, alluding to predictability (e.g., Keenlyside et al. 2008; Meehl et al. 2009; 2014). However the proxy reconstruction and CMIP5 analysis showed that even though both Atlantic SSTs and ISM have persistent multi-decadal variability, their relation is not stable. The CMIP5
analysis further showed that external forcing, including volcanic eruptions, might be the reason for their apparent relation, abating the possibility for prediction of the ISM on multi-decadal timescales. To complicate the picture further AMV can also modulate interannual variability in the tropics, and while coupled GCMs are able to simulate ENSO to some degree the tropical Atlantic has larger biases, inhibiting the ability to simulate Atlantic Niños (Richter and Xie 2008) and consequently the relation between Atlantic and Pacific Niños. Since decadal variability in the ISM is a modulation of interannual variability, it is not surprising that coupled GCMs have difficulties simulating this. However state-of-the-art coupled models do simulate a variety of AMV amplitudes and frequencies, and the simulated teleconnection patterns might depend on biases in the North Atlantic as well (Wang et al. 2014). Understanding these issues could ultimately lead to skillful decadal predictions of regional climate variability in the Indo-Pacific and over the Northern Hemisphere.
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Paper I

Marine-based multiproxy reconstruction of Atlantic multidecadal variability

Marine-based multiproxy reconstruction of Atlantic multidecadal variability

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Abstract

Atlantic multidecadal variability (AMV) is known to impact climate globally, and knowledge about the persistence of AMV is important for understanding past and future climate variability, as well as modeling and assessing climate impacts. The short observational data do not significantly resolve multidecadal variability, but recent paleoproxy reconstructions show multidecadal variability in North Atlantic temperature prior to the instrumental record. However, most of these reconstructions are land-based, not necessarily representing sea surface temperature. Proxy records are also subject to dating errors and microenvironmental effects. We extend the record of AMV 90 years past the instrumental record using principle component analysis of five marine-based proxy records to identify the leading mode of variability. The first principal component is consistent with the observed AMV, and multidecadal variability seems to persist prior to the instrumental record. Thus, we demonstrate that reconstructions of past Atlantic low-frequency variability can be improved by combining marine-based proxies.

1. Introduction

In this study we reconstruct the Atlantic multidecadal variability (AMV) by combining several marine-based proxy records from the North Atlantic region. During the instrumental period, North Atlantic sea surface temperature (SST) has undergone pronounced basin-wide fluctuations, with warm and cold periods of 3-4 decades each. These variations are referred to as AMV or the Atlantic Multidecadal Oscillation [e.g., Enfield et al., 2001]. Studies suggest that the AMV is important for climate variability globally and has been connected to several regional climate signals. The AMV can, for instance, affect European and North American climate [e.g., Sutton and Hodson, 2005; Wyatt et al., 2012], the frequency of Atlantic hurricanes [e.g., Goldenberg et al., 2001], and Arctic sea ice cover [e.g., Kinnard et al., 2011; Miles et al., 2014]. It has also been linked to changes in rainfall in the African Sahel [e.g., Zhang and Delworth, 2006; Wang et al., 2012], as well as the South Asian summer monsoon [e.g., Goswami et al., 2006].

Whether the AMV is a persistent mode of internal variability is still disputed [Kilbourne et al., 2008; Knudsen et al., 2011]. Ocean temperature data are limited to the last 140 years by instrumental records, and the data are spatiotemporally lacking before 1950 [Smith and Reynolds, 2003]. The relatively short instrumental SST record can therefore only capture 1–2 cycles of AMV, and is too short to confidently study natural low-frequency variability.

High-resolution climate reconstructions based on long-lived marine biota, for instance, tropical corals, bivalve mollusks, and coraline algae, can help reconstruct SST prior to the instrumental era [Jones et al., 2001; Wannamaker et al., 2011; Hetzinger et al., 2012]. With such reconstructions we can investigate whether multidecadal variability of Atlantic SST is a persistent feature of the Atlantic climate. Alternative tools for studying AMV are climate models. However, state-of-the-art climate models simulate a wide range of variability because of large uncertainties in the underlying processes [Medhaug and Furevik, 2011; Ba et al., 2014].

While here we reconstruct the AMV using marine-based proxies, previous reconstructions of the AMV have mainly used land-based proxies, such as tree rings [e.g., Gray et al., 2004; Mann et al., 2009]. Many of these land-based reconstructions have also used records from regions far from the Atlantic Ocean, in noncoastal areas. How the low-frequency variability in SST is related to atmospheric temperatures is, however, not clear. The relationship between SST and tree ring proxies seems strong for the instrumental era, but this relationship may not be stable [D’Arrigo et al., 2008; Vásquez-Bedoya et al., 2012]. Therefore, we investigate...
low-frequency variability of North Atlantic SST using marine-based proxies, as these are a more direct measure of SST compared to land-based proxies.

There are several annual-resolution marine-based proxy records that capture the AMV signal [e.g., Hetzinger et al., 2008; Kilbourne et al., 2008; Saenger et al., 2009; Halfar et al., 2011]. Many of these proxy records are relatively short, not extending past the instrumental era, but their consistency with the instrumental record indicates that these types of records can be used to reconstruct past SST. However, individual records will be subject to sampling and dating errors as well as microenvironmental effects. Principle component analysis (PCA) can be used to reduce these uncertainties by extracting the leading patterns of variability [Storch and Zwiers, 2002]. While this method is common for land-based proxies, its application to marine-based proxies has been limited [Ault et al., 2009]. Here we successfully apply this method to reconstruct AMV 90 years past the instrumental data period.

2. Data and Method

We analyze five published marine-based proxy records from the North Atlantic sector (Table 1). These records are chosen on the basis that they all have annual resolution, are longer than the instrumental record, and are proxies for SST. The proxy records stem from massive-growing tropical coral colonies. Two of the records are based on growth [Saenger et al., 2009; Vásquez-Bedoya et al., 2012], two are based on the Sr/Ca ratios [Goodkin et al., 2005; Kilbourne et al., 2008], and one is based on the skeletal δ¹⁸O composition [Swart et al., 1996]. The locations of these proxy records are illustrated in Figure 1a, and their properties are summarized in Table S1. The records used in the analysis are all from the western tropical Atlantic. Cooler temperatures in the eastern Atlantic lead to slower coral growth rates, hampering the formation of large and long-lived colonies of massive growing coral. Thus, no long-term coral-based proxies exist at the moment from the eastern Atlantic, as well as at higher latitudes.

The five coral records from the tropical Atlantic are combined with PCA, where the individual records have been detrended and normalized prior to the analysis. Combining these records gives us an overlapping time span of 206 years from 1781 to 1986. For validation of our AMV reconstruction we compare our results with the SST data from Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST) [Rayner et al., 2003]. We also compare our results with two other AMV reconstructions [Gray et al., 2004; Mann et al., 2009]. Both these records are multiproxy reconstructions, but neither focuses on marine proxies. The reconstruction from Gray et al. [2004] is composed of 12 tree ring chronologies located in North America, Europe, and the Middle East between 30°N and 70°N [Gray et al., 2004]. Although the reconstruction from Mann et al. [2009], which is composed of several different types of records, includes some marine records from the North Atlantic, none of them has continuous annual resolution [Mann et al., 2009].

3. Results and Discussion

The first principal component (PC1) from the PCA explains 32% of the variance in the records and exhibits a close correspondence to the observed AMV-index (Figure 1b). The AMV-index is defined as the detrended and normalized annual averaged low-frequency (11 year running mean) Atlantic SST averaged over the region 0–60°N and 75°W–7.5°W [Enfield et al., 2001; Wyatt et al., 2012]. The time series of PC1 is consistent with

<table>
<thead>
<tr>
<th>Table 1.</th>
<th>Loadings on PC1 for the Proxy Records and the Correlation With the AMV-Index*</th>
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<tbody>
<tr>
<td>Reference</td>
<td>Loadings on PC1</td>
</tr>
<tr>
<td>Goodkin et al. [2005]</td>
<td>0.48</td>
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<tr>
<td>Kilbourne et al. [2008]</td>
<td>0.05</td>
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<tr>
<td>Saenger et al. [2009]</td>
<td>0.45</td>
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<tr>
<td>Swart et al. [1996]</td>
<td>-0.28</td>
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<tr>
<td>Vásquez-Bedoya et al. [2012]</td>
<td>0.70</td>
</tr>
<tr>
<td>Composite</td>
<td>0.52</td>
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<tr>
<td>AMV from Gray et al. [2004]</td>
<td>0.57</td>
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<tr>
<td>AMV from Mann et al. [2009]</td>
<td>0.57</td>
</tr>
<tr>
<td>PC1</td>
<td>0.53</td>
</tr>
</tbody>
</table>

*aThe correlation with the AMV-index is calculated for the included proxy records, the composite of the proxy records, the AMV reconstruction from Gray et al. [2004] and Mann et al. [2009], and PC1, for the period 1871–1986.
the instrumental SST record from the Atlantic with warm periods from 1860 to 1890 and 1940 to 1970, and a cold period from 1900 to 1930. The time series have a correlation of 0.53 at a zero lag. This and other correlations discussed below are summarized in Table 1. PC1 captures an additional cold period in the 1930s that is not present in the observed AMV-index. This cooling is present in four out of the five proxy records included in our analysis, suggesting that this cold period is a tropical signal. However, this cooling is not present in the observed records for the western tropical Atlantic. Prior to the instrumental record, PC1 has a cold period from 1820 to 1860 and a warm period from the beginning of the record (1781) to 1820. When we repeat the PCA for the coral records without detrending the records first, PC1, now explaining 35% of the variability, captures the same periodicity and timing of warm and cold periods as for the detrended records (not shown). In addition, there is a positive linear trend. PC1 provides evidence that the multidecadal variability in Atlantic SSTs may have persisted prior to the instrumental record.

PC1 is also comparable to the AMV reconstruction from both Gray et al. [2004] (G04) and Mann et al. [2009] (M09), but the reconstructions are somewhat displaced in time (Figure 2). For instance, the first warm period in the beginning of PC1 ends later than the corresponding warm period in the land-based reconstructions; hence, the cold period that lasts until about 1860 starts later in PC1 (Figure 1b). The cross correlation between PC1 and the AMV reconstruction from Gray et al. [2004] has a maximum value of 0.44 when PC1 lags the reconstruction by 11 years. At a zero lag the correlation is 0.26. The cross correlation between PC1
and the AMV reconstruction from Mann et al. [2009] has a maximum value of 0.39 when PC1 lags the reconstruction by 12 years. At a zero lag the correlation between PC1 and the reconstruction from Mann et al. [2009] is 0.33, slightly higher than that for the reconstruction from Gray et al. [2004]. The lags are similar when the reconstructions are 10 year low-pass filtered, as seen in Figure 2.

A significant positive correlation between PC1 and the observed SST is found for the whole tropical North Atlantic (Figure 1a). The same correlation pattern is found for PC1 with the records not detrended prior to the analysis (not shown). The highest correlations are found in the tropics in the region where the corals are situated. This bias in the location of the coral records may distort our results, with PC1 capturing a tropical rather than an extratropical North Atlantic signal. The correlation of PC1 with observed North Atlantic SSTs averaged over the tropics (0°–30°N, 75°W–7.5°W) and the sub tropics (30°–60°N, 75°W–7.5°W) are 0.59 and 0.47, respectively. The correlation of PC1 with observed North Atlantic SSTs averaged over western (0°–60°N, 75°W–45°W) and eastern (0°–60°N, 45°W–7.5°W) North Atlantic are 0.53 and 0.52, respectively. The correlation between the 10 year low-pass filtered PC1 and observed SSTs show a similar pattern; however, the correlations are lower (see Figure S1 in the supporting information). We do not see such a strong tropical bias in the correlation pattern between the other two land-based multiproxy AMV reconstructions and observed Atlantic SST (Figure S1). For the observed SSTs, correlations are also high in the tropics; however, there are also high correlations in the sub tropics that are not visible in the correlation with PC1 (Figure S2). The observed SST pattern is also subject to uncertainty [Alexander et al., 2013], and in particular the patterns for the period prior to and following the 1940s differ markedly (Figure S2). Interestingly, the pattern prior to the 1940s resembles that associated with PC1.

Correlations between the observed AMV-index from HadISST with each individual proxy record, the composite of the records, the two land-based AMV reconstructions [Gray et al., 2004; Mann et al., 2009], and PC1 is given in Table 1. The correlation with PC1 is 0.57 and is higher than the correlation with any of the individual proxy records alone, except for the record from Vásquez-Bedoya et al. [2012]. This gives indications that PC1 is an improved reconstruction of AMV compared to individual proxy records. The correlations for the observed AMV-index with the two AMV reconstructions from Gray et al. [2004] and Mann et al. [2009] are slightly higher than for the correlation with PC1. This might be because these reconstructions have been calibrated to the observed SST field [Gray et al., 2004; Mann et al., 2009]. In addition, the 12 tree ring chronologies used in the reconstruction from Gray et al. [2004] have been chosen due to their strong link to Atlantic SSTs in the observational record.

PC1 is the optimally weighted average of the records, with loadings given in Table 1, while a composite is a simple average of the proxy records. Thus, PCA extracts the common variability in the proxy records, and can act as a filter to eliminate higher frequencies of variability. The correlation of the observed AMV-index with PC1 is equal to the correlation of the observed index with the composite (unweighted average) of the proxy records, and PC1 is not necessarily an improved AMV reconstruction compared to the composite. Nevertheless, PCA may give more reliable results than an unweighted average because PCA allows some records to be lightly weighted or even negatively weighted.

When working with proxy records, there is always the possibility that the proxy may not actually be reflecting the variable of interest. In our study we are interested in temperature, but the marine-based proxies may also be influenced by, for instance, salinity. However, we find that our reconstruction is able to successfully reproduce the variability seen in the instrumental SST record, and also by combining several records using PCA we may be able to extract the SST signal as a dominant mode of low-frequency variability.

### 4. Conclusion

Here we propose a different method for reconstructing low-frequency variability in North Atlantic SST based on records from long-lived marine biota. We combine several annual resolution marine-based proxy records, extending 90 years further back in time than the instrumental record, with PCA to extract the low-frequency variability and limit microenvironmental effects and sampling errors. We find that PC1 is consistent with the observed AMV, and that the AMV persists throughout the record. This suggests that our method is able to capture the Atlantic low-frequency variability, and we conclude that this method is adequate for reconstructing SST on multidecadal timescales.
We find a discrepancy in the timing of the variability between our marine-based reconstruction and other land-based multiproxy AMV reconstructions, with the land-based AMV reconstructions leading 11–12 years to the marine-based reconstruction, indicating that we have to be careful about using proxies for reconstructing multidecadal SST variability in the Atlantic. There are also discrepancies related to the location of the proxies used. For our marine-based reconstruction we find higher correlations in the tropics and an additional cold period in the late 1930s. These discrepancies reflect differences in using proxies from high and low latitudes and of various types, as well as errors and uncertainties in the records. However, there are at present relatively few high-resolution long marine proxy records from the Atlantic sector, and the existing records are mainly from the tropics. Longer marine records, including records from the subtropics, are needed to reconstruct the AMV even further back in time. Additional high-resolution marine-based proxy records will also improve the confidence in AMV reconstructions, and will help constrain climate models and in turn predictions. Such reconstructions can also be used to investigate the persistence of observed AMV teleconnections. This record is freely available for use.

Acknowledgments
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References


### Table S1. Proxies used in the analysis

<table>
<thead>
<tr>
<th>Reference</th>
<th>Location</th>
<th>Species</th>
<th>Proxy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Goodkin et al., 2005</td>
<td>32°N, 64°W</td>
<td>Diploria labyrinthiformis</td>
<td>Sr/Ca</td>
</tr>
<tr>
<td>Kilbourne et al., 2008</td>
<td>17°N, 67°W</td>
<td>Montastraea faveolata</td>
<td>Sr/Ca</td>
</tr>
<tr>
<td>Saenger et al., 2009</td>
<td>26°N, 79°W</td>
<td>Siderastrea siderea</td>
<td>growth</td>
</tr>
<tr>
<td>Swart et al., 1996</td>
<td>25°N, 80°W</td>
<td>Montastraea faveolata</td>
<td>δ¹⁸O</td>
</tr>
<tr>
<td>Vásquez-Bedoya et al., 2012</td>
<td>21°N, 87°W</td>
<td>Siderastrea siderea</td>
<td>growth</td>
</tr>
</tbody>
</table>
Figure S1. Correlation between the annually averaged observed SST (HadISST) and the decadal filtered (a) PC1 and AMV-reconstructions from (b) Gray et al. [2004] and (c) Mann et al. [2009] for the period 1871-1986.
Figure S2. Correlation between the observed AMV-index and annually averaged observed SST (HadISST) for the period (a) 1871-1986, (b) 1871-1940 and (c) 1941-2012, showing only values at a 95% confidence level for the effective degrees of freedom.
Paper II

Teleconnections between Indian summer monsoon rainfall and Atlantic multidecadal variability over the last 500 years

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Manuscript in preparation
Teleconnections between Indian summer monsoon rainfall and Atlantic multidecadal variability over the last 500 years

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Abstract

Several studies have linked Atlantic multidecadal variability (AMV) with Indian summer monsoon rainfall (ISMR). Instrumental records of both ISMR and North Atlantic sea surface temperatures (SSTs) have multidecadal variability with a period close to 60 years, with periods of warm (cold) North Atlantic SSTs accompanied by periods of wetter (drier) ISMR and lower (higher) frequencies of dry monsoon years. In this paper we have studied both AMV and ISMR for the period from 1481 to present using several proxy reconstructions with annual resolution from both regions, to investigate multidecadal variability in ISMR and the link to AMV prior to the instrumental period. We find that multidecadal variability in the ISMR is persistent, but the link between ISMR and AMV may not be persistent. The observed correlation between AMV and ISMR has also weakened in the last decade, further suggesting that the AMV-ISMIR link is not stable.

1. Introduction

Multidecadal variability has been observed in the Indian summer monsoon rainfall (ISMR) (e.g., Parthasarathy et al., 1994). The four-month period from June through September constitutes the Indian summer monsoon season, and gives about 78% of the mean annual rainfall of India (Mooley and Parthasarathy, 1984), with a standard deviation of about 10% of the mean. Parthasarathy et al. (1994) found that the measured rainfall data available from 1871 to present showed decadal departures of ISMR above and below the long-term average for 30-year epochs. The years 1901-1930, and 1961-1990 were epochs with low decadal mean monsoon rainfall and frequent dry years; whereas 1871-1900 and 1931-1960 were epochs with high decadal
mean monsoon rainfall and infrequent dry years. This approximate 60-year period is found to be in phase with Atlantic multidecadal variability (AMV) (Figure 1; e.g., Goswami et al., 2006). The AMV is associated with basin-wide low-frequency sea surface temperature (SST) variability in the North Atlantic (e.g., Enfield et al., 2001), and has earlier been related to decadal variability in a number of regions, such as European and North American summer climate (e.g., Sutton and Hodson, 2005) and rainfall in African Sahel, as well as ISMR (e.g., Zhang and Delworth, 2006). A significant correlation between the observed AMV and ISMR suggests a link between these two regions. Here we investigate if this link holds before the instrumental period using several available annual resolution proxy reconstructions.

Several studies have investigated the teleconnection between AMV and ISMR based on observational data and modeling experiments (e.g., Goswami et al., 2006; Zhang and Delworth, 2006; Luo et al., 2011; Li et al., 2008). Goswami et al. (2006) showed that the AMV produced persistent weakening (strengthening) of the meridional gradient of tropospheric temperature by setting up a negative (positive) tropospheric temperature anomaly over Eurasia during boreal summer/autumn resulting in an early (late) withdrawal of the monsoon and a persistent decrease (increase) of seasonal monsoon rainfall. Zhang and Delworth (2006) suggested that during the positive AMV phase the Intertropical Convergence Zone (ITCZ) over the Atlantic shift northwards resulting in anomalous south westerly surface winds over India, convergence of surface moisture and thus enhanced summer monsoon rainfall over India. The positive AMV phase reduces sea level pressure over Asia and Europe and snow cover over Tibet, which in turn strengthens the monsoon rainfall over India. Luo et al. (2011) found that the influence of AMV on ISMR was achieved through an atmospheric teleconnection process in which a propagating Rossby wave train from the North Atlantic across South Asia enhanced the South Asia high. In general studies have found that a positive AMV favors greater than normal monsoon rainfall over India. However this relation is not robust in coupled model simulations (Ting et al., 2011), and it is not clear whether this link on decadal timescales is causal and internal to the climate system, or specific for the observational time period.
A connection between ISMR and North Atlantic SSTs has also been indicated on centennial and millennial time-scales, linking cold events in the North Atlantic to a weakening of the Asian monsoon during the Holocene and the last glacial period (Gupta et al., 2003; Burns et al., 2003). Similarly Overpeck et al. (1996) found this linkage to be significant for glacial inter-glacial cycles. Co-variations between ice-core and tree-ring derived surface temperatures of the Tibetan Plateau (TP), the ISM and North Atlantic SSTs have also been found on interglacial timescales in proxy records covering the past 2000 years (Feng and Hu, 2005; 2008). The proxy data used in these studies mostly have decadal-to-centennial or lower temporal resolution.

The aim of our study is to investigate the relationship between decadal variability of ISMR and AMV for the past 500 years, from 1481. Therefore we have collected and analyzed the available ISMR data using a combination of high-resolution (near annual) proxy records and rain gauge observations, and compared with available AMV reconstructions. In Section 2 we analyze the available instrumental data of ISMR and AMV. Section 3 describes the proxy data used to estimate ISMR and AMV in this study. Decadal variability in the proxy records is assessed in Sections 4 and compared in Section 5. These results are discussed and summarized in Section 6.

2. Instrumental data of ISMR and AMV

Our period of investigation, 1481-2010, can be divided into two parts: 1844-2010 when instrumental rainfall data from India are available and 1481-1843 when proxy data are available. Previous studies investigating the relation between AMV and ISMR in instrumental data have only used the period from 1871 and onwards, but rain gauge data are available from 1813 (Sontakke et al. 1993). Since the distribution of the stations is quite sparse from 1813 to 1843, only rain gauge data from the year 1844 are included here, extending the instrumental record by 26 years. Starting with 10 rain gauge stations in 1844 the number increased progressively to 19 by year 1870 (Figure 2a). In this study we define years with ISMR less (more) than one standard deviation of the long-term mean as dry (wet) years. We analyze decadal variability, but interannual variability is important as we study decadal variability of the frequency of
dry monsoon years. Epochs of frequent dry years are defined as a DRY epoch, and epochs of infrequent dry years are defined as a WET epoch (Joseph 1976). Based on this definition, we characterized the period 1844 to 1860 as a DRY epoch, with four dry years and no wet years, while the decade 1861-1870 had above average rainfall (Table 1).

Parthasarathy et al. (1994) constructed an ISMR time series from 1871 using rain gauge data from 306 stations well distributed over India (Mooley and Parthasarathy, 1984; Figure 2b), annually updated by the Indian Institute for Tropical Meteorology (Parthasarathy et al., 1995). Table 1 shows the dry years identified in this data together with the data spanning years 1844-1870. The previously mentioned 30-year DRY and WET epochs have occurred alternatively since the 1840s. The 30-year epoch 1991-2020 is therefore expected to be a WET epoch, and has currently had only 5 dry years, the last one in 2015.

There are several ways to define the AMV index, but all definitions are based on North Atlantic SST anomalies, low-frequency filtered or smoothed to capture multidecadal variability (e.g., Enfield et al., 2001). For the instrumental period we have used the un-smoothed version of the NOAA PSD AMV index starting from year 1856, which is the detrended area-weighted average of the Kaplan Extended SST V2 data set (Kaplan et al., 1998) over the North Atlantic (0 to 70° N). We have then averaged the index over the monsoon months (June to September), and smoothed using an 11-year moving average.

3. Available proxy data and selection
Since reliable instrumental data only go back about 150 years, we need to use proxy reconstructions to study decadal variations in both ISMR and North Atlantic SSTs prior to this period. Here we present previously published AMV and ISMR reconstructions with annual resolution, using proxy records from tree rings, speleothems and corals, in order to critically select the data for our analysis.
3.1. Proxy records of AMV

There are presently three available annual resolution multi-proxy AMV reconstructions: the reconstruction from Gray et al. (2004), from Mann et al. (2009) and from Svendsen et al. (2014) (Table 2). The record from Gray et al. (2004) is based on 12 tree ring reconstructions from around the North Atlantic and spans the years 1567-1990. Mann et al. (2009) used a global proxy dataset comprising thousands of tree ring, ice core, coral, sediment, and other proxy records spanning the ocean and land regions globally over the past 1500 years to reconstruct surface temperature changes on a global scale. In this study we have used the AMV data reconstructed by Mann et al. (2009) for the period that overlaps with the ISMR reconstructions, years 1481-2006. The record from Svendsen et al. (2014) is reconstructed using five marine-based proxy records from massive-growing tropical coral colonies from the North Atlantic and covers the years 1781-1986. These three proxy reconstructions of AMV have been validated earlier, and they compare well in the instrumental period (Svendsen et al., 2014), however discrepancies between the records are present before this period. At the moment we do not know which of these reconstructions are more reliable, and more high-resolution proxy reconstructions are needed to assess these discrepancies (Kilbourne et al., 2014; Svendsen et al., 2014).

3.2. Proxy records of ISMR

In the tropics high precipitation periods favor growth in tree rings, while periods of drought inhibit their growth (Borgaonkar et al., 1994, 1996, Pant et al., 1988, 1998, Yadav et al., 1999) Speleothem records from caves yield similar rainfall estimates as tree rings (Yadava and Ramesh, 1999a; 1999b; Yadava et al., 2004; Yadava and Ramesh, 2005; Sinha et al., 2007). Corals can also be used as proxies for the conditions that they live in (Druffel, 1997). By assuming a general relation between ISMR and Indo-Pacific SSTs associated with the El Niño-Southern Oscillation (ENSO) (Rasmusson and Carpenter, 1983; Krishnamurthy and Goswami, 2000), ISMR can be reconstructed based on annual banding or chemical composition of coral skeletons from the Pacific.
Only five published high-resolution ISMR proxy reconstructions were available to us, and are presented here: two based on tree rings, the Kerala Tree Ring Chronology (KTRC; Borgaonkar et al., 2010) and the Statistical Model Monsoon Rainfall (SMMR; Pant et al., 1988), two speleothem records from the Dandak cave (Sinha et al., 2007) and the Jhumar cave (Sinha et al., 2011), and one coral record from Palmyra Island in the equatorial Pacific (Chakraborty et al., 2012). All these reconstructions have near-annual or higher resolution and have been studied in relation to ISMR (Table 3).

The Palmyra Coral record is short and does not overlap in time with the instrumental record for ISMR and we can therefore not evaluate if this record is able to capture the dry and wet years in ISMR. Both speleothem records are from the monsoon core region where local precipitation is strongly correlated with the instrumental ISMR-index (Gadgil, 2003), and therefore have ideal locations for ISMR reconstructions. However, the millennial scale Jhumar cave record with an average temporal resolution of 1.45 years does not capture the dry years in the instrumental record (not shown). The Dandak Cave record, as for the Palmyra coral record, is for a period that is not comparable with the instrumental data and hence cannot be evaluated. We therefore chose to continue our analysis using only the two tree ring reconstructions, KTRC and SMMR, both with annual resolution.

The KTRC is a combined tree-ring width chronology prepared from three sites (shown in italics in Figure 2a) along the Western Ghats mountains in the state of Kerala for overlapping periods: Tekkedy for period 1785-2003, Narangathara for period 1742-2003 and Nellikooth for period 1481-2003 (Borgaonkar et al., 2010). KTRC has been validated with the instrumental data and showed that low tree growth is associated with deficient ISMR since the late 19th century. Before that period several low tree growth years coincide with El Niño years often consistent with dry ISMR years, or with Indian droughts in historical records. Periods with high growth rate may or may not represent wet periods, as excess moisture does not necessarily result in higher tree growth for this data (Borgaonkar et al., 2010). The KTRC is therefore only a time series indicating dry years, but does not indicate the magnitude of these anomalies.
Most of the dry years determined from the instrumental rainfall data are identified in KTRC with an accuracy of +/-2 years.

The SMMR is based on a reconstruction of the Southern Oscillation using tree ring chronologies from both western North America and the Southern Hemisphere for the period 1600-1961 (Lough and Fritts, 1985). Pant et al. (1988) used this data to reconstruct the ISMR over the period 1602-1960, hereinafter SMMR, based on the known relation between the Southern Oscillation and ISMR. The mean rainfall of ~850 mm in SMMR is comparable to the instrumental ISMR data for the period 1871-1960, but the estimated standard deviation is quite low, at 35 mm compared to 81 mm for the instrumental data for the same period (Figure 3). The correlation coefficient between instrumental ISMR and SMMR is 0.43 for this overlapping period, statistically significant at a 99% confidence level. During the 90-year period 1871-1960 that overlaps the instrumental period, 15 dry years were identified in the SMMR as opposed to 11 in the instrumental record, with a hit rate of 73% (11 out of 15) indicating that SMMR is effective in identifying dry rainfall years.

The two ISMR proxy records and the instrumental ISMR record are compared with the India Meteorological Department daily (IMD) gridded rainfall data, available from 1901 (Rajeevan et al. 2005) in Figure 4 and 5. The correlation between KTRC and the mean June to September IMD gridded rainfall data for their overlapping period is shown in Figure 4a and the correlation between instrumental ISMR and IMD gridded rainfall data is shown in Figure 4b. The correlation with KTRC and instrumental ISMR data show similar spatial pattern with significant correlations over north-west and central India and low correlations over southern Peninsular India and north eastern regions. The corresponding correlations between SMMR and IMD gridded rainfall data for their shorter overlapping period are shown in Figure 5. The spatial pattern for SMMR is similar to that of KTRC, but correlations are weaker.

4. Evaluation of decadal variability in ISMR prior to the instrumental period
Decadal variability in North Atlantic SST anomalies before the instrumental period
has been evaluated in earlier studies (Gray et al., 2004; Mann et al., 2009; Svendsen et al., 2014). In this section we evaluate the decadal variability of ISMR in the two proxy reconstructions chosen in Section 3, the KTRC and SMMR.

In KTRC we find, as in Borgaonkar et al. (2010), that there are several lengthy low growth periods, indicating decadal variability in ISMR. We have grouped the KTRC into seven temporal clusters of frequent dry years with a total of 86 dry years (Table 4 and Figure 6). The total duration of these DRY epochs (A1-G1) are 282 years, with an average frequency of one dry year in about three years. The periods between the DRY epochs, which may be called WET epochs, had only 6 dry years (the years 1581, 1594, 1600, 1724, 1735 and 1778) in a total duration of 228 years, with a frequency of one dry year in 38 years. These frequencies compare well with the frequencies of dry years in DRY and WET epochs in the instrumental period. The time difference in years between the middle points of the successive DRY epochs are 55, 123, 86, 70, 73 and 57 years which may be taken as the period of the decadal variation. The period is in the range of 55 to 86 years, comparable to the decadal variability of both AMV and ISMR in the instrumental records, except for one long 123-year dry spell. To quantify the wet and dry years, since the KTRC only gives us an estimate of the timing of dry years, we construct a rainfall time series based on KTRC where dry years are represented by value 1 and non-dry years by 0. An 11-year moving average of this rainfall series divided by 11 to get the probability of a dry year is shown in Figure 6, with the seven DRY epochs labeled. The probability of a dry year reaches up to 40-50 percent during some periods. The timing of the DRY epochs in KTRC is comparable to the observed DRY epochs. For instance years containing the epoch E1 is partially covered by the observations (starting from 1844) and the number of observed dry years during the period 1844 to 1877 is comparable to the KTRC data. However, the number of dry years in the earlier half of the 20th century (1900-1930) is overestimated by KTRC (epoch F1), whereas the number of dry years in the latter half of the 20th century (after 1970) is underestimated (epoch G1).

In the SMMR as for the instrumental ISMR data, we define those years with rainfall
less than one standard deviation from the long-term mean as a dry year. Table 5 gives
a complete list of the dry years and indicates clusters of frequent dry years in the
SMMR reconstruction for the period 1602-1960 (Table 5 and Figure 6). The
alternating 30-year DRY and WET epochs seen in the instrumental data is also
discernible in SMMR. The last cluster of dry years from 1897 to 1930 (epoch E2)
coincides well with the DRY epoch in the instrumental ISMR of 1899 to 1920 (see
Figure 6). The cluster of dry years in SMMR from 1676 to 1720 (epoch A2) covers a
major part of dry epoch C1 in KTRC, 1743 to 1776 (epoch B2) includes dry epoch D1
and period 1790 to 1811 (epoch C2) and 1863 to 1876 (D2) are part of dry epoch E1 in
KTRC. The five dry epochs given by SMMR has a total duration of 149 years and a
total number of 41 dry years, with an average frequency of one dry year per 3.6 years.
This compares well with the frequency of dry years in DRY epochs of ISMR in the
instrumental rainfall data.

5. Relation between ISMR and AMV for the proxy period
As seen in previous studies, the 30-year DRY and WET epochs in the instrumental
ISMR time series coincide with the cold and warm epochs in observed AMV (Fig.1).
Earlier studies have also shown that AMV has persistence prior to the instrumental
records, and the section above suggests that multidecadal variability has persistence in
the ISMR as well. Here we compare the two selected ISMR proxy reconstructions,
SMMR and KTRC, with the three available AMV reconstructions, Gray-AMV, Mann-
AMV and Svendsen-AMV, to investigate the stability of the observed AMV-ISMR
relation. We have used the negative of the KTRC record since dry years are defined as
1 and other years as 0. Such a quantitative comparison of annual resolution ISMR
proxy records has never been published, and in light of recent modeling studies on the
AMV-ISMR link, a comparison with available AMV reconstructions is overdue. In the
following analysis we have used an 11-year moving average filter on all the time
series. We have also tested our results with other more advanced filters, and the results
are similar. The published AMV index from Mann et al. (2009) is already low-pass
filtered, and this data is therefore not additionally smoothed for our analysis.
The instrumental and proxy reconstructed AMV and ISMR indices have a common overlapping period of 131 years (1856 to 1986), except for SMMR that only covers the years up to 1960. The linear correlation coefficients between all these indices are estimated for this common period and for the entire length of the two indices being compared (Table 6). For the common period there are significant correlations between all the AMV proxy data with the observed AMV, with the Gray-AMV having the highest correlation of 0.83. Both proxy reconstructions of ISMR have a significant positive correlation with the instrumental ISMR index.

The correlation between observed AMV and ISMR for the period 1854-2010 is 0.43, but this strengthens to 0.64 when we only take into account the shorter overlapping common period 1856-1986. The correlation seems sensitive to end points, but the observed AMV and ISMR have diverged since the 1990s (Figure 1). The observed AMV has significant positive correlations with both ISMR proxy records. The correlation with SMMR is higher, but the overlapping period is also shorter since SMMR only extends to year 1960. However when we calculate the correlations between the AMV and ISMR proxy records for the entire length of the records, the correlation between AMV and ISMR breaks down (Table 6). There are no significant correlations between the proxy reconstructions of AMV and ISMR, except between Svendsen-AMV and SMMR, which has the shortest overlapping period. Both ISMR proxy records, KTRC and SMMR, even show negative correlation with Mann-AMV for the whole overlapping periods. While the proxy reconstructions for both AMV and ISMR have positive correlations with the observed records, the positive AMV-ISM correlation does not persist throughout the time period studied here back to 1481.

The filtered time series of observed and proxy reconstructed ISMR and AMV over the entire period of their availability are shown in Figure 7. ISMR and AMV are more or less in phase during the 20th century, except for at the very end of the century. We can also see that the DRY ISMR epoch identified in the earlier instrumental records from 1844-1860 correspond to a cold period in the reconstructed AMV indices. The AMV proxy data are in good agreement until the middle of the 18th century (Figure 7b-d),
but they deviate considerably during the decades prior to this, consistent with the correlations in Table 6. The ISMR proxy indices KTRC and SMMR are in close agreement with each other during the entire time period available for comparison (1602-1960), except during the 18th century when the two proxy data differ in phase over a number of decades. It should be kept in mind that KTRC only gives a qualitative measure of the dry years and does not distinguish between excess and normal rainfall year, possibly leading to some of these phasing differences. The discrepancies between the proxy reconstructions emphasize the importance of using several reconstructions when comparing proxies.

6. Discussion and Conclusions

Decadal variability of ISMR for the recent five centuries has been studied here using instrumental rain gauge data and available high-resolution proxy reconstructions. From the period of instrumental data (1844-2010) we found that the alternate 30-year epochs of frequent and infrequent dry years extend back to 1844. The epochs 1844-1870, 1900-1930 and 1961-1990 had frequent dry years with an average of one dry year in 3 years, and the decadal mean ISMR was below the long-time average. The 30-year epochs 1871-1900 and 1931-1960 had very few dry years and the decadal mean ISMR was above the long-time average. The period 1484 to 1843 had several DRY epochs with about one dry year in 3 years on average, and the decadal variability had a period of 55 to 86 years. Thus during the entire period 1484 to 2010 a prominent multidecadal signal is present in the ISMR.

The link between AMV and ISMR during the period 1481-2010 was analyzed by comparing ISMR reconstructions with instrumental data and available proxy reconstructions of AMV. For the period when instrumental data was available for comparison, we see that during the WET (DRY) epochs of ISMR the North Atlantic had above (below) average SSTs. However our analyses showed that the observed positive correlation between the AMV and ISMR during the instrumental period is not seen prior to 1750, although multidecadal variability is present in the proxy reconstructions of both ISMR and AMV. We also notice that the observed AMV-
ISMIR link has weakened in the last decade, suggesting that the positive correlation between these two regions is not stable.

We have validated the data sets used, and as concluded in Section 3 we found that the two ISMR reconstructions we have used, KTRC and SMMR, were the most ideal for this analysis. However there are drawbacks in these data sets also and proxy reconstructions have uncertainties related to sampling and dating errors. For instance, the KTRC data set is based on tree rings from only one region in India, the state of Kerala in south-west India. Although the correlation between all-India monsoon rainfall and rainfall in the state of Kerala is statistically significant, the record is likely to include regional rainfall variability. However, this record seems to capture the majority of the instrumental ISMR dry years. SMMR is based on a reconstruction of the Southern Oscillation, and takes advantage of the observed relation between ENSO and ISMR. Even though the relation between ISMR and ENSO has been strong in the instrumental period there are studies indicating that this relation has broken down since the 1990s (e.g., Krishna Kumar et al., 1999). The SMMR reconstruction depends on a stable ENSO-ISMR relation, which might not be realistic. In addition to the uncertainties in the ISMR proxy reconstructions, the AMV reconstructions also have discrepancies prior to year 1750. Despite these limitations, our results suggest that the observed AMV-ISMR link is not persistent. The weakening correlation in the recent decade further supports these results. However due to inconsistencies in available proxy records for both AMV and ISM, more reconstructions are needed to further confirm or refute the results presented here.

Earlier studies have identified a robust relation between North Atlantic SST and the Asian monsoon on millennial time-scales using low-resolution proxy data (e.g., Gupta et al., 2003, Jung et al., 2004). The results found here are not in disagreement with this as our focus has been on shorter timescales of variability using annual resolution proxy data. SST anomalies in the Atlantic approximated for millennial time-scale variations including periods of major climate shifts are larger than those characterizing the AMV during the past 500 years, and could therefore also have larger impacts.
To summarize, we find that the observed correlation between AMV and ISMR extends back to the 18th century. However none of the records show consistent positive correlations prior to that period, indicating that the AMV-ISMR link may not be persistent. The weakening observed in the latest decade suggest that other physical mechanisms for driving multidecadal variability in the ISMR may be important, for instance the SST gradient between the tropics and Northern midlatitudes suggested by Joseph et al. (2013). However uncertainties in the proxy reconstructions contribute to the insignificant AMV-ISMR correlation before the 18th century, and the results imply that better quality data is required to be conclusive, and caution should be exercised when using single proxy reconstructions for comparing with records from other regions as well as with model simulations.

Acknowledgements
We thank Dr. G. B. Pant and co-authors for supplying the rainfall data derived from global tree rings using the statistical method, and Dr. H. P. Borgaonkar for useful discussions regarding the Kerala Tree Ring series. We are grateful to Prof. Noel Keenlysde for helping us improve the manuscript. This work was done as part of the India-Clim project (no. 216554) funded by Research Council of Norway. G. Bindu was funded by the EU INDO-MARECLIM project coordinated by Nansen Environmental Research Centre India (NERCI).

References
L.F.) trees from south India. Palaeogeogr Palaeoclimatol Palaeoecol 285(1-2), 74-84.


Monsoon over the last 18000 years. Climate Dynamics 12, 213-225.


Table 1 Dry and wet years of ISMR for the period 1844 - 2010

<table>
<thead>
<tr>
<th>Decade</th>
<th>Decadal mean</th>
<th>Dry years</th>
<th>Wet years</th>
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<tbody>
<tr>
<td>1844-1850</td>
<td>824.4</td>
<td>1848</td>
<td>-</td>
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<tr>
<td>1851-1860</td>
<td>829.6</td>
<td>1851, 1855, 1860</td>
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<td>1861-1870</td>
<td>895.0</td>
<td>1864</td>
<td>1861, 1862, 1863, 1867, 1870</td>
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<td>1871-1880</td>
<td>849.9</td>
<td>1873, 1877</td>
<td>1874, 1878</td>
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<tr>
<td>1881-1890</td>
<td>881.4</td>
<td></td>
<td>1884</td>
</tr>
<tr>
<td>1891-1900</td>
<td>865.5</td>
<td>1899</td>
<td>1892, 1893, 1894</td>
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<td>1901-1910</td>
<td>822.4</td>
<td>1901, 1904, 1905</td>
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<td>1911-1920</td>
<td>821.3</td>
<td>1911, 1918, 1920</td>
<td>1916, 1917</td>
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<tr>
<td>1921-1930</td>
<td>837.2</td>
<td>-</td>
<td>-</td>
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<tr>
<td>1931-1940</td>
<td>871.3</td>
<td>-</td>
<td>1933</td>
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<td>1941-1950</td>
<td>888.8</td>
<td>1941</td>
<td>1942, 1947</td>
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<tr>
<td>1951-1960</td>
<td>871.4</td>
<td>1951</td>
<td>1956, 1959</td>
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Table 2 Available AMV proxy reconstructions used in this study

<table>
<thead>
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<th>Name</th>
<th>Reference</th>
<th>Type</th>
<th>Period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gray-AMV</td>
<td>Gray et al., 2004</td>
<td>Tree ring</td>
<td>1567-1990</td>
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<tr>
<td>Mann-AMV</td>
<td>Mann et al., 2009</td>
<td>Multi-proxy</td>
<td>1481-2006</td>
</tr>
<tr>
<td>Svendsen-AMV</td>
<td>Svendsen et al., 2014</td>
<td>Coral</td>
<td>1781-1986</td>
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Table 3 Available ISMR proxy reconstructions

<table>
<thead>
<tr>
<th>Name</th>
<th>Type</th>
<th>Period</th>
<th>Region</th>
<th>Resolution</th>
<th>Reference</th>
</tr>
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<tbody>
<tr>
<td>KTRC</td>
<td>Tree ring</td>
<td>1481-2003</td>
<td>Kerala, India</td>
<td>annual</td>
<td>Borgaonkar et al., 2010</td>
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<tr>
<td>SMMR</td>
<td>Tree ring</td>
<td>1602-1960</td>
<td>Global</td>
<td>annual</td>
<td>Pant et al., 1988</td>
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<tr>
<td>Dandak</td>
<td>Speleothem</td>
<td>600-1500</td>
<td>Central India</td>
<td>~annual</td>
<td>Sinha et al., 2007</td>
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<td>Jhumar</td>
<td>Speleothem</td>
<td>1075-2008</td>
<td>Central India</td>
<td>~annual</td>
<td>Sinha et al., 2011</td>
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<tr>
<td>Palmyra</td>
<td>Coral</td>
<td>1635-1702</td>
<td>Pacific</td>
<td>seasonal</td>
<td>Chakraborty et al., 2012</td>
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Table 4 Epochs of frequent dry years from 1481-2003 derived from KTRC

<table>
<thead>
<tr>
<th>DRY epoch</th>
<th>Dry years</th>
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<tr>
<td>(A1)</td>
<td>1484,1486,1491,1492,1508,1510,1512,1514,1515</td>
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<td>(B1)</td>
<td>1539,1540,1541,1552,1564,1565,1569,1570,1571</td>
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<tr>
<td>(C1)</td>
<td>1652,1656,1663,1665,1666,1676,1680,1681,1691,1693,1694,1696,1700,1703</td>
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<tr>
<td>(D1)</td>
<td>1760,1761,1763,1765,1767</td>
</tr>
<tr>
<td>(E1)</td>
<td>1791,1794,1802,1803,1811,1812,1814,1818,1821,1824,1829,1832,1833,1837,1838,1844,1853,1865,1868,1873,1877</td>
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<tr>
<td>(F1)</td>
<td>1896,1899,1901,1904,1905,1908,1911,1914,1915,1918</td>
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</table>

Table 5 Epochs of frequent dry years from 1602-1960 derived from SMMR

<table>
<thead>
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<th>DRY epoch</th>
<th>Dry years</th>
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<tr>
<td>(A2)</td>
<td>1676, 1679, 1682, 1688, 1700, 1702, 1705, 1717, 1720</td>
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<tr>
<td>(B2)</td>
<td>1743, 1746, 1747, 1754, 1755, 1758, 1760, 1768, 1769, 1775, 1776</td>
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<tr>
<td>(C2)</td>
<td>1790, 1792, 1799, 1807, 1808, 1811</td>
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<tr>
<td>(D2)</td>
<td>1863, 1864, 1868, 1876</td>
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<tr>
<td>(E2)</td>
<td>1897, 1899, 1902, 1904, 1911, 1912, 1913, 1914, 1918, 1925, 1930</td>
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</table>
Table 6 Linear Correlation Coefficients between AMV and ISMR indices normalized and smoothed with an 11-year moving average for the period 1856-1986. The correlations for the entire overlapping record length and the years covered are given in parenthesis.

<table>
<thead>
<tr>
<th></th>
<th>Gray AMV</th>
<th>Mann AMV</th>
<th>Svendsen AMV</th>
<th>Observed ISMR</th>
<th>KTRC</th>
<th>SMMR#</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Observed AMV</strong></td>
<td>0.83*</td>
<td>0.55***</td>
<td>0.52***</td>
<td>0.64*</td>
<td>0.35</td>
<td>0.52***</td>
</tr>
<tr>
<td></td>
<td>(1856-1990: 0.83*)</td>
<td>(1856-2006: 0.54***</td>
<td>(1856-1986: 0.52***</td>
<td>(1856-2010: 0.43***</td>
<td>(1844-2000: 0.33)</td>
<td>(1856-1960: 0.52***</td>
</tr>
<tr>
<td><strong>Gray AMV</strong></td>
<td>0.78*</td>
<td>0.42</td>
<td>0.55**</td>
<td>0.26</td>
<td>0.27</td>
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</tr>
<tr>
<td></td>
<td>(1567-1990: 0.50*)</td>
<td>(1781-1986: 0.33)</td>
<td>(1844-1990: 0.51**</td>
<td>(1567-1990: 0.06)</td>
<td>(1602-1960: -0.03)</td>
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</tr>
<tr>
<td><strong>Mann AMV</strong></td>
<td>0.39</td>
<td>0.30</td>
<td>0.07</td>
<td>0.22</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>(1781-1986: 0.47***</td>
<td>(1844-2006: 0.03)</td>
<td>(1491-2000: -0.07)</td>
<td>(1602-1960: -0.19)</td>
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<tr>
<td><strong>Svendsen AMV</strong></td>
<td>0.49***</td>
<td>0.16</td>
<td>0.83*</td>
<td></td>
<td>0.56***</td>
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<td></td>
<td>(1844-1986: 0.48**)</td>
<td>(1781-1986: 0.10)</td>
<td>(1781-1960: 0.46**)</td>
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<td>(1844-1960: 0.45***</td>
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<tr>
<td><strong>Observed ISMR</strong></td>
<td>0.42</td>
<td></td>
<td></td>
<td>0.57**</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>(1844-2000: 0.32)</td>
<td></td>
<td></td>
<td>(1602-1960: 0.41**)</td>
<td></td>
<td></td>
</tr>
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</table>

SMMR# - Linear correlation coefficient is calculated for the period 1856-1960.

* indicates a confidence interval of 99% for the effective degrees of freedom.

** indicates a confidence interval of 95% for the effective degrees of freedom.

*** indicates a confidence interval of 90% for the effective degrees of freedom.
Figure 1 Instrumental ISMR (red) and the AMV-index (blue) from June to September smoothed with an 11-year moving average (1856-2010).

Figure 2 (left) Locations of the 19 selected stations from Sontakke et al. (2003). Mountainous regions (hatched) were not considered for the study. The stations are AMT (Amraoti), BMB (Bombay), BNG (Bangalore), BRL (Bareilly), DSA (Deesa), DLH (Delhi), FZP (Ferozepur), GRK (Gorakhpur), HYD (Hyderabad), JBP (Jabalpur), MDS (Madras), NGP (Nagpur), PTN (Patna), PNE (Pune), PRI (Puri), RPR (Raipur), SMG (Shimoga), SNI (Seoni) and VNS (Varanasi). The location of the three tree ring sites used in KTRC is shown in italics. (right) Network of 306 raingauge stations over the area considered excluding hilly area (hatched) from Parthasarathy et al. (1994).
Figure 3 SMMR (red line) and instrumental ISMR (blue) for the period 1871-1960.

Figure 4 (left) Correlation between KTRC and IMD gridded rainfall data and (right) observed ISMR and IMD gridded rainfall data for the period 1901-2000. Dry years in KTRC are represented with 1 and non-dry years with 0. Only correlations significant at 95% confidence level are shown.

Figure 5 (left) Correlation between SMMR and IMD gridded rainfall data and (right) observed ISMR and IMD gridded rainfall data for the period 1901-1960. Only correlations significant at 95% confidence level are shown.
Figure 6 11-year moving average of the ISMR reconstructions based on KTRC (black), SMMR (green) and instrumental ISMR (red) where dry years are represented by 1 and non-dry years by 0. The number of dry years per successive 11-year period divided by 11 (probability of drought) is given by the vertical axis. A1-G1 and A2-E2 mark the DRY epochs in KTRC (Table 4) and SMMR (Table 5), respectively.
**Figure 7** (a) Observed AMV (blue) and ISMR (red) and (b-d) proxy reconstructed ISMR, KTRC (yellow) and SMMR (orange), and (b) Gray-AMV, (c) Mann-AMV and (d) Svendsen-AMV (blue). All the time series data are normalized and smoothed using an 11-year moving average.
Paper III

External forcing synchronizes Atlantic multidecadal variability and the Indian summer monsoon

Luo F., Y. Gao, L. Svendsen, N. Keenlyside, S. Li and T. Furevik

Manuscript in preparation
External forcing synchronizes Atlantic multidecadal variability and the Indian summer monsoon

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\textsuperscript{1}Nansen-Zhu International Research Centre and Climate Change Research Center, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China, \textsuperscript{2}Nansen Environmental and Remote Sensing Center and Bjerknes Centre for Climate Research, Bergen, Norway, \textsuperscript{3}Geophysical Institute, University of Bergen and Bjerknes Centre for Climate Research, Bergen, Norway

Abstract

The relationship between Atlantic multidecadal variability (AMV) and the Indian summer monsoon (ISM) is analyzed in five models from the CMIP5 that have higher skills in simulating both AMV and ISM among 25 models. The significant AMV-ISM relation in observations is not simulated by any of the five models in pre-industrial control simulations with fixed external forcing, but is captured by one model (GFDL-CM3) in the historical simulations including transient external forcings. Further analyses of GFDL-CM3 reveal that external forcing is linked to a strengthened land-ocean thermal gradient over South Asia consistent with an enhanced ISM, which also leads to a concurrent evolution of AMV. Thus the significant relationship found in observations may be associated with external forcing, instead of internal climate variability.

1. Introduction

The Atlantic multidecadal variability/oscillation (AMV/AMO) is a leading pattern of variation in sea surface temperatures (SSTs) in the North Atlantic. The AMO has been considered to be internal climate variability associated with the Atlantic meridional overturning circulation (AMOC) (e.g., Delworth and Mann, 2000; Zhang and Delworth, 2005; Knight et al., 2005; Msadek et al., 2011; Medhaug and Furevik, 2011; Drinkwater et al., 2014; McCarthy et al., 2015). The AMV is traditionally regarded as a synonym for the AMO, but is more relevant for the observed variations as we cannot
be sure of its oscillatory nature (Drinkwater et al., 2015). Many regional climate variations have been associated with the AMV, such as North American and European summer climate (Sutton and Hodson, 2005, 2007), Atlantic hurricanes (e.g., Zhang and Delworth, 2006), Sahel summer rainfall (Zhang and Delworth, 2006; Ting et al., 2011), the East Asian monsoon (Li and Bates, 2007; Lu et al., 2006; Wang et al., 2009; Zhou et al., 2015), and the Indian summer Monsoon (ISM) (e.g., Goswami et al., 2006; Li et al., 2008; Luo et al., 2011). However, the AMV and its associated impacts might also be influenced by solar, volcanic, and aerosol forcings and stratosphere circulation anomalies (Otterå et al., 2010; Booth et al., 2012; Reichler et al., 2012; Chiang et al., 2013; Dunstone et al., 2013; Cheng et al., 2013; Wilcox et al., 2013; Martin et al., 2014; Allen, 2015).

Earlier observational and modeling studies (e.g., Goswami et al., 2006; Zhang and Delworth, 2006; Feng and Hu, 2008; Li et al., 2008; Luo et al., 2011) reveal that a positive (negative) AMV phase corresponds to a strengthening (weakening) of the ISM. The physical mechanisms of the AMV impact on the ISM can be classified into two types: through air-sea interactions where a large reduction in the Atlantic thermohaline circulation moves the intertropical convergence zone (ITCZ) southwards, weakens the surface trade winds over the Pacific, induces an El Nino-like/La Nina-like pattern and a weakened/enhanced Walker circulation in the southern/northern tropical Pacific, and weakens the ISM (Zhang and Delworth, 2005; Stager et al., 2011), or through a more direct atmospheric teleconnections where a positive AMV causes a tropospheric warming over Europe and western Asia, excites a Rossby wave train from the North Atlantic across South Asia, and increases the meridional temperature gradient between Eurasia and the Indian Ocean and thus increases the ISM rainfall (Li et al., 2008; Wang et al., 2009; Luo et al., 2011). However, Lu et al. (2006) found that a warm phase of the AMV caused a late withdrawal of the ISM instead of enhanced ISM rainfall. Ting et al. (2011), using 23 CMIP3 models, found no connection between the AMV and the ISM in pre-industrial simulations, but a positive connection in historical simulations although the signal is 2
weak. It should be mentioned that most of the earlier studies addressing the relation between AMV and the ISM have focused on either the post-industrial era (after 1860) (e.g., Li et al., 2008; Wang et al., 2009; Lu et al., 2006) or assuming that the North Atlantic is forced by a strong increase in freshwater (e.g., Zhang and Delworth, 2005; Yu et al., 2010; Stager et al., 2011). However Luo et al. (2011) did find a positive AMV-ISM connection in a 600-year pre-industrial simulation using a single model. Thus it remains unclear whether the AMV-ISM connection seen in observations is an innate feature of the climate system, externally forced, or rather an artifact of the short observational record. Here, we explore whether the relationship may result from internal climate variability or from external forcing by comparing pre-industrial simulations with historical 20th century simulations from multiple CMIP5 models.

The paper is organized as follows. Section 2 describes the method and data sets used in this study. Section 3 analyses the observed and modeled data, and investigates the relationship between the AMV and the ISM. Finally, a discussion and summary are given in section 4 and 5.

2. Dataset and Method
We analyzed 25 models participating in the CMIP5 and found that five of the models better capture the temporal and spatial characteristics of the observed AMV and ISM (see Auxiliary Material Section S1 for details). These five models are used to further investigate the AMV-ISM relationship (Table 1). Two types of experiments are utilized: (1) “Pre-industrial Control” simulations (here labeled CTL) with a length of minimum 400 years (Table 1), forced by non-evolving pre-industrial conditions, including prescribed well-mixed gases, natural aerosols or their precursors, and some short-lived species; solar forcing is kept constant and there are no volcanoes. CTL only has one ensemble member for each model. (2) “Historical” simulations for the period 1860-2005 (labeled ALL), with observed forcing agents, including emissions or concentrations of well-mixed greenhouse gases, natural and anthropogenic aerosols, solar forcing and land-use changes (Taylor et al., 2012). The ensemble means of each
model are calculated, and the numbers of ensemble members are listed in Table 1. Moreover, three additional sets of historical simulations of the GFDL-CM3 are used: (1) a three-member ensemble with the natural forcing agents only (referred to as NAT), (2) a three-member ensemble with the anthropogenic forcing agents only (referred to as ANT), and (3) a three-member ensemble with the anthropogenic aerosols only (referred to as AA). Modeled monthly global SST, surface temperature, precipitation and air temperature are used, downloaded from the project for model diagnostics and intercomparisons (PCMDI) CMIP5 website (http://pcmdi9.llnl.gov/esgf-web-fe/) and the IPCC Data Distribution Centre (http://www.ipcc-data.org/).

The SST dataset used is the Met Office Hadley Centre’s monthly SST records (HadISST) (Rayner et al., 2003) spanning the years 1870–2010 and gridded to 1.0° latitude by 1.0° longitude. In addition, we use the ISM rainfall from 1871–2010 based on more than 300 stations around India (Rajeevan et al., 2006). For investigating the possible impacts of the external forcing, a volcanic aerosol forcing time series derived from ice core peaks in sulphate/conductivity has been used (Crowley et al., 2003), as well as a solar irradiance time series (Bard et al., 2000).

The AMV-index is here defined as the annual averaged SST anomalies in the North Atlantic (0°-60°N, 75°W-7.5°W) (Enfield et al., 2001; Sutton and Hodson, 2005), and the ISM-index as the seasonal (June-September) averaged land rainfall over India (10°N-30°N, 60°E-90°E) (Goswami et al., 2006; Li et al., 2008). The global mean surface temperature (GMT) index is defined as the annual surface temperature anomalies averaged over the globe from 60°N to 60°S. All of the indices are filtered with a 9-year running mean to obtain the low-frequency components. The observed datasets and the output from the historical experiments are further linearly detrended to reduce the global warming signal. To account for serial correlation, the degrees of
freedom for statistical tests are estimated by \( N_\varepsilon = \frac{N}{1 + 2 \sum_{i=1}^{10} a_i b_i} \), where \( N \) is the number of years, and \( a_i \) and \( b_i \) are the \( i \)th order autocorrelation for time series \( a \) and \( b \), respectively.

3. Results

As many earlier studies suggested (e.g., Goswami et al., 2006; Zhang and Delworth, 2006; Feng and Hu, 2008; Li et al., 2008; Wang et al., 2009), there is a significant positive correlation (\( R=0.5 \), \( P<0.05 \)) between the observed AMV and the ISM for the period from 1874 to 2001. The two cold AMV phases from the early 1900s to the late 1920s and from the mid-1960s to the 1990s correspond to ISM dry phases, and the two warm phases correspond to wet ISM phases (Figure S1).

However, the significant observed correlation cannot be found in the five selected models in CTL (Table 2). This indicates that in these models internal climate dynamics are not able to reproduce the observed link between AMV and the ISM. For ALL, one model (GFDL-CM3) yields a significant positive correlation (\( R=0.69 \) for five ensemble members) stronger than in the observed records, while the other four models show insignificant correlations. The correlation range for the different ensemble members of GFDL-CM3 is from 0.31 to 0.54 (Table S1). The correlations are outside the range found in a sliding correlation with a 128-year time window (a window with the length of the instrumental data and ALL simulations) between the AMV-index and the ISM-index for GFDL-CM3 in CTL, showing no periods in CTL when the correlation is as high as in the historical ensemble. The maximum correlation in CTL for GFDL-CM3 is 0.31 (Figure S2). The relatively high correlation in ALL in GFDL-CM3 seems to be associated with external forcing and not due to random fluctuations. The sliding correlations for the other 4 models show that the correlation is never as high as in observations (Figure S2). The signal to noise ratio in GFDL-CM3 is 1.64 for the AMV-indices, and 0.56 for the ISM-indices. Here we define the signal to noise
ratio as the variance of the ensemble mean over the total variance. It is worthwhile to mention that the correlation of the modeled ISM-index with the observed is relatively low \(R=0.15\), while the correlation between the AMV-indices is significant \(R=0.63\), suggesting that external forcing has a stronger influence on the AMV (Otterå et al., 2010) than on the ISM.

Considering that only GFDL-CM3 of the five selected models simulates the observed AMV-ISM relation, we employed GFDL-CM3 to investigate how the external forcing could impact this connection (hereafter CTL and ALL refer only to GFDL-CM3). Figure 1 shows the regression patterns of surface temperature onto the ISM and the AMV indices in GFDL-CM3 for CTL and ALL. In CTL, the wet ISM phase is weakly associated with a cold Pacific Decadal Oscillation (PDO)-like spatial pattern over the North Pacific and a tripole pattern in the North Atlantic (Figure 1a). For the warm AMV phase, there is a basin-scale warming with a maximum located south of Greenland over the North Atlantic, and slight warming over the North Pacific (Figure 1b). There is little similarity in the North Atlantic between the two regression patterns in CTL, with a pattern correlation of \(R=-0.13\) over the area from \(0^\circ-60^\circ\)N and \(75^\circ\text{W}-7.5^\circ\text{W}\). In contrast both the ISM and AMV in ALL are related to a similar global warming pattern, and the warming amplitudes in the Northern Hemisphere extratropics are larger than in the tropics (Figure 1c, 1d). This appears consistent with Joseph et al. (2013), who found that multidecadal variability of the ISM has a higher correlation with the Northern Hemisphere meridional SST gradient between the tropics and the extra-tropics than with the AMV-index.

Several previous studies (e.g., Goswami et al., 2006; Wang et al., 2009; Luo et al., 2011) suggest that modifying the meridional gradient of the tropospheric temperatures (MGT) between the Indian subcontinent and the tropical Indian Ocean is the primary mechanism for the AMV impact on the ISM. A MGT-index (MGTI) is defined as the difference in vertically averaged temperature between a northern box \((10^\circ\text{N}-35^\circ\text{N}, 30^\circ\text{E}-100^\circ\text{E})\) and a southern box \((15^\circ\text{S}-10^\circ\text{N}, 30^\circ\text{E}-100^\circ\text{E})\) (Goswami et al., 2006;
Luo et al., 2011). For the positive ISM phase, there is a band of warming from northern Africa to India and no significant anomalies over the tropical Indian Ocean, corresponding to an enhanced land-ocean MGT (MGTI: 0.07°C) (Figure 1e). In contrast, during a positive AMV a weak warming is found over a band from 30°S to 60°N, with a negative land-ocean thermal gradient (MGTI: −0.006°C) in the ISM sector (Figure 1f), consistent with the insignificant relation between AMV and the ISM in CTL. In ALL, positive AMV and ISM periods exhibit increased 500-200hPa temperature globally, with maximum warming located at the east coast of North America and from the north of Africa to the south of Japan (Figure 1g, 1h). Hence, positive phases in both the ISM and AMV are associated with a positive MGT between the tropical Indian Ocean and the Indian subcontinent (MGTI: 0.03°C for ISM and 0.004°C for AMV).

Basically, in CTL the ISM is closely related to the land-ocean thermal gradient between Eurasia and the Indian Ocean (Figure 1e) and also partly to the Atlantic tripole pattern and the PDO-like pattern (Figure 1a). In ALL the ISM is also related to the land-ocean thermal gradient with warming north of India (Figure 1g), in addition to a wide spread warming over the extra-tropics (Figure 1c), partly in agreement with the study of Kumar et al. (1999). Similarly, in ALL the AMV relation to surface and 500-200hPa temperature is more global compared to CTL, and a positive AMV is related to an increased land-ocean thermal contrast in the Indian Ocean sector (Figure 1d), which is weaker in CTL (Figure 1b). Hence, in ALL both the positive ISM and AMV are associated with the positive land-ocean MGT between Eurasia and the Indian Ocean.

In CTL the ISM has no statistical relation to the GMT and AMV with correlation coefficients of -0.09 and 0.02, respectively (Figure 2a). The relation between the GMT and the AMV in CTL is stronger with a correlation coefficient of 0.61. The
correlations among GMT, AMV and ISM indices are significantly strengthened in ALL relative to in CTL with correlation coefficients of 0.95 between GMT and AMV, 0.66 between GMT and ISM, and 0.69 between AMV and ISM (Figure 2b). The coherence between the three variables is consistent with all three being driven by an external factor. Comparing the regression patterns of vertical 500-200hPa mean temperature onto the GMT-index in CTL and ALL (Figure 2c-d), shows that the maximum heating area shifts northwards in ALL, and three maximum centers are located over the Gulf Stream region, eastern Mediterranean and eastern Asia, respectively. This pattern is similar to the regression patterns with the ISM and AMV indices in ALL (Figure 1g, 1h). The heating center over the Gulf Stream region could be linked to atmospheric circulation changes over the North Atlantic that may contribute to drive the AMV (further analysis is beyond the scope of this study). While this simple analysis comparing CTL and ALL suggests that external forcing drives the enhanced land-ocean MGT and may explain the observed relation between AMV and the ISM during the instrumental period, the mechanisms remain unclear.

4. Discussion
We use the GFDL-CM3 simulations to further explore the relative effects of natural forcing (NAT), anthropogenic forcing (ANT) and anthropogenic aerosols (AA) on the AMV-ISM relationship (Figure 3). There is a positive correlation between AMV and the ISM in NAT (R=0.43) (Figure 3a). Although not significant at the 5% level, the correlation strengthens relative to the correlation of 0.02 in CTL, indicating that natural forcing factors might play a role in modulating the correlation. A lead-lag correlation analysis in ALL shows that there is no correlation between the solar radiation time series (Bard et al. 2000) and the ISM or the AMV. For volcanoes, a superposed epoch analysis (Haurwitz and Brier, 1981) using the volcanic aerosol forcing time series from Crowley et al. (2003) shows that maximum negative responses in both the AMV and the ISM indices are found five years after strong volcanic explosions (Table S2). It can also be seen in Figure 3a that the declines of the two indices correspond to large volcanic explosions. Volcanic aerosols have been
linked to the phasing of the AMV before the 20th century as well (Otterå et al., 2010). The connection between AMV and ISM in ANT and AA is weak (R=0.13 and R=−0.19, respectively) (Figure 3b, 3c). This analysis implies that volcanic aerosols could be the main cause for modulating the relationship between AMV and the ISM. But it should be noted that the analysis above is based on three-member ensemble means and a larger ensemble is necessary to better remove the influence of internal variability in the analysis. However Booth et al. (2012) suggested that aerosol forcing through aerosol-cloud microphysical effects is a prime driver of the AMV in the historical period. Similarly Bollasina et al. (2011) used the same model as we have, GFDL-CM3, to link anthropogenic aerosol forcing to ISM changes since the 1950s on decadal timescales. This illustrates that aerosol forcing, including both volcanic and anthropogenic, may synchronize the connection between AMV and the ISM. Considering that only GFDL-CM3 of the five selected models simulates a significant correlation between AMV and ISM, we speculate the aerosol indirect effects, which are only included in GFDL-CM3, play a key role in this correlation.

Compared to a previous modeling study using a pre-industrial multi-century simulation with the Bergen Climate Model (BCM) (Luo et al. 2011) that suggested that the AMV can impact the ISM through an atmospheric teleconnection pattern internal to the climate system, this study gives an alternative hypothesis with the AMV-ISM relation being externally forced, and shows that the BCM model response is an exception compared to other CMIP5 models (Ting et al., 2011; Kavvada et al., 2013).

5. Summary

Given the uncertain relationship between the AMV and the ISM in previous studies (e.g., Goswami et al., 2006; Zhang and Delworth, 2006; Lu et al., 2006; Li et al., 2008; Ting et al., 2011), we have analyzed CMIP5 simulations to better understand this relation, and how and why it may have changed over time. CMIP5 simulations provide
evidence suggesting that the recent AMV-ISM co-variation in observations may not be intrinsic to the climate system, and the significant positive correlation in GFDL-CM3 in historical runs may result from external forcing. Comparing pre-industrial runs (CTL) and historical runs (ALL) in GFDL-CM3, shows that the ISM and AMV are only significantly correlated in ALL, both associated with the land-ocean thermal gradient between Eurasia and the Indian Ocean. Analysis indicates that the mid-to-upper tropospheric temperature changes over southern Eurasia could be driven by external forcing. These changes can directly affect the ISM, while external forcing may also drive the AMV by influencing the local atmospheric circulation in addition to direct thermal forcing.

To summarize, we found: (1) the significant AMV-ISM connection is not present in the control simulations of the five selected CMIP5 models, and is therefore possibly not a feature of the internal climate system; (2) the significant AMV-ISM connection after 1860 could be caused by external forcing modulating both the AMV and ISM.

Acknowledgments
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References


Feng, S., and Q. Hu (2008), How the North Atlantic Multidecadal Oscillation may have influenced the Indian summer monsoon during the past two millennia,


Martin, E., C. Thornicroft, and B. B. B. Booth (2014), The Multidecadal Atlantic SST-


### Table 1 Details of the five CMIP5 models used in this study

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<tr>
<th>Modeling group</th>
<th>Model</th>
<th>Ocean (Y, X)</th>
<th>Atmosphere (Lon×Lat)</th>
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<th>ALL (ensemble size)</th>
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### Table 2 The correlation coefficients between AMV and ISM in CTL and ALL. Significant values at the 5% level are in bold.

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Figure 1  Regressions of filtered surface temperature (Ts) and vertical mean air temperature between 500 hPa and 200 hPa ($T_{500-200}$) onto the standardized ISM and AMV indices in (a-b and e-f) CTL and (c-d and g-h) ALL for the GFDL-CM3 model. Black dots indicate significance levels exceeding 5% according to a t-test. The black boxes in (e) indicate the northern box (10°N-35°N, 30°E-100°E) and the southern box (15°S-10°N, 30°E-100°E) used to define the MTGI. Units: °C.
Figure 2 (a-b) Temporal evolution of the GMT, AMV and ISM indices in CTL and ALL in GFDL-CM3. (c-d) Regressions of vertical mean air temperature between 500 hPa and 200 hPa onto the standardized GMT indices in (c) CTL and (d) ALL for the GFDL-CM3 model. Black dots indicate significance levels exceeding 5% according to a t-test. Units: °C
Figure 3 Temporal evolution of the AMV and ISM indices for (a) NAT, (b) ANT and (c) AA in GFDL-CM3. The radiative forcing from volcanic eruptions is shown in (a) (vertical black lines).
Auxiliary Material

**Table S1** The correlation between AMV and ISM indices for the different ensemble members in GFDL-CM3

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**Table S2** Superposed epoch analysis for the ISM and AMV indices.

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**Figure S1** Temporal evolution of the observed AMV and ISM indices.
Figure S2 Sliding correlations (window: 128 years) between the AMV-index and the ISM-index in the 5 selected CMIP5 models. The straight horizontal line indicates the observed correlation coefficient (R=0.5).
Section S1: Comparison of CMIP5 models

The skills of 25 CMIP5 models in simulating the observed AMV and ISM are assessed here, based on outputs from historical experiments (Table SS1). We finally selected 5 models (BCC-CSM1.1(m), CNRM-CM5, GFDL-CM3, MPI-ESM-LR, and MPI-ESM-P) with higher simulated skills to investigate the connection between the AMV and the ISM in this study.

1.1 The AMV

First we looked at the temporal characteristics, including period and amplitude. The AMV index is defined as the annual averaged SST anomalies in the North Atlantic basin (0°-60°N, 75°W-7.5°W) (Enfield et al., 2001). All of the indices are filtered with a 9-year running mean filter and linearly detrended. Spectral analysis shows that the majority of the models simulate significant decadal-to-multidecadal variability in the 10-100 year timescale, except FGOALS-g2 and GFDL-ESM2G (Figure SS1). Then via the Taylor diagram (Figure SS2) (Taylor et al., 2001), it can be seen that 4 models have amplitudes outside ±0.5 standard deviation of observations.

The modeled spatial features are compared with observations in Figure SS3 and Figure SS4. The simulated skills of different models vary greatly, representing by the wide range of spatial correlation coefficients (SCCs) (from -0.3 to 0.6) and standard deviations (Figure SS4). The SCCs in 5 models are less than or equal to zero (Figure SS4). The standard deviations in 9 models are greater than one standard deviation of the observations (Figure SS4), which is likely due to the large amplitude around Greenland in these models (Figure SS3). Table SS2 summarizes the above evaluation criteria and shows that there are 10 models that better capture the observed spatiotemporal features of the AMV.

The summer precipitation associated with the AMV over the tropical Atlantic and Sahel is evaluated in the 10 remaining models, since many studies have demonstrated the robust relationships between a positive AMV, a northward shift of the Atlantic
ITCZ and more Sahel rainfall (e.g. Zhang et al., 2006; Ting et al., 2011). 6 models exhibit the feature of the northward shift of the Atlantic ITCZ as well as increased rainfall on the northern side of the climatological ITCZ and decreased rainfall on the southern side (Figure SS5), and more Sahel rainfall can be seen in all models (Figure SS5).

1.2 The ISM

All of the 6 remaining models (BCC-CSM1.1(m), CNRM-CM5, GFDL-CM3, HadGEM2-ES, MPI-ESM-LR, and MPI-ESM-P) capture well the observed seasonality of the ISM including the onset in June (Figure SS6). The amplitudes of 4 models are weaker than the observed, while the remaining 2 are comparable to (GFDL-CM3) or much stronger (BCC-CSM1.1(m)) than observation (Figure SS6). The rainfall pattern in all of the models resemble the observed climatological ISM rainfall (June-September) with a maximum in the west of India (Fig. SS7). However HadGEM2-ES does not simulate the location of the observed ITCZ in the Indian Ocean leading to a low SCC of 0.56 with observations (Figure SS7e). The remaining 5 models show better performance with higher SCCs, greater than 0.6 (Figure SS7).

References


Table SS1 Details of the 25 CMIP5 models evaluated in this study

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<th>Ref. #</th>
<th>Model ID</th>
<th>Modeling group</th>
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Table SS2 The evaluation criteria of AMV for the models. A cross indicates that: (1) the models have no significant multidecadal period (3rd column); (2) the temporal standard deviations are outside the range of ±0.5 standard deviation of the observations (4th column); (3) the spatial correlations between observations and the models are negative (5th column); (4) the spatial standard deviations are outside the range of ±1 standard deviation of the observations (6th column).

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Figure SS1 Power spectrum of the AMV-indices for (0) observations (HadISST) and (1-25) the models. The power spectrum is given by the black line, significant above the red line at the 5% level.

Figure SS2 Taylor diagram for the temporal features of the AMV-indices for the period 1874-2001. The x-axis shows normalized standard deviations and the arc shows the correlation values between observations and the ensemble mean for each model. The red number represents the reference number for each model listed in Table SS1.
Figure SS3 Regressions of SST onto the AMV-indices in (0) observation and (1-25) models. Black dots indicate significance levels exceeding 95% according to a t-test. Units: °C.
**Figure SS4** Same as Figure SS2 but for the spatial features of the regressions displayed in Figure SS3. Note that the model (GFDL-ESM2G) has a normalized standard deviation larger than 4.0 and is not shown.
Figure SS5 Regressions of filtered (9-year running mean) precipitation onto the standardized AMV-indices in the models [Units: mm/day]. The black contours are for climatological precipitation contoured at 2 mm/day. The green boxes represent the Sahel region (10°N–20°N, 20°W–40°E).
Figure SS6 Seasonal evolution of the ISM-index for observations (CMAP; solid line) and the models (dashed lines). The ISM-index is defined as the averaged land rainfall over India (10°N—30°N, 60°E—90°E). Unit: mm/day.
Figure SS7  Climatological precipitation for (a) observations (CMAP) and (b-g) models. Values in parentheses indicate the spatial correlation coefficients of the models with the observation. Unit: mm/day
Paper IV

Weakening AMOC connects equatorial Atlantic and Pacific interannual variability

Svendsen, L., N. G. Kvamstø, and N. Keenlyside (2013)

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Weakening AMOC connects Equatorial Atlantic and Pacific interannual variability

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Abstract Observations indicate that since the 1970s Equatorial Atlantic sea surface temperature (SST) variations in boreal summer tend to modulate El Niño in the following seasons, indicating that the Atlantic Ocean can have importance for predicting the El Niño–Southern Oscillation (ENSO). The cause of the change in the recent decades remains unknown. Here we show that in the Bergen Climate Model (BCM), a freshwater forced weakening of the Atlantic meridional overturning circulation (AMOC) results in a strengthening of the relation between the Atlantic and the Pacific similar to that observed since the 1970s. During the weakening AMOC phase, SST and precipitation increase in the central Equatorial Atlantic, while the mean state of the Pacific does not change significantly. In the Equatorial Atlantic the SST variability has also increased, with a peak in variability in boreal summer. In addition, the characteristic timescales of ENSO variability is shifted towards higher frequencies. The BCM version used here is flux-adjusted, and hence Atlantic variability is realistic in contrast to in many other models. These results indicate that in the BCM a weakening AMOC can change the mean background state of the Tropical Atlantic surface conditions, enhancing Equatorial Atlantic variability, and resulting in a stronger relationship between the Tropical Atlantic and Pacific Oceans. This in turn alters the variability in the Pacific.

Keywords El Niño–Southern Oscillation · Atlantic Zonal Mode · Atlantic Niño · Atlantic Meridional Overturning Circulation · Bergen Climate Model · Interannual variability

1 Introduction

The El Niño–Southern Oscillation (ENSO) is the dominant mode of natural climate variability on interannual time-scales and one of the most important sources of variability in the Tropical Pacific (e.g., Philander 1990; Latif and Keenlyside 2009). ENSO-like variability also exists in the Tropical Atlantic and is commonly referred to as the Atlantic Zonal Mode or Atlantic Niño (e.g., Zebiak 1993; Latif and Grötzner 2000; Keenlyside and Latif 2007; Jansen et al. 2009). However, variability in the Atlantic is weaker than in the Pacific and it peaks in boreal summer, rather than in boreal winter (Keenlyside and Latif 2007). In this study, we investigate the relationship between the Pacific and the Atlantic Niños.

ENSO can affect climate, ecology and economy in the Tropical Pacific region, as well as weather and climate globally (Brönnimann 2007). An example of such widespread effects is the warm phase of ENSO in the winter of...
1997/1998 that contributed to a high global surface temperature record in 1998 (Latif and Keenlyside 2009). There have been numerous studies on how the global climate is affected by ENSO, for instance the effect of ENSO on the extratropics (Wallace and Gutzler 1980; Hoskins and Karoly 1981), and on remote regions such as the Atlantic (e.g., Zebiak 1993; Enfield and Mayer 1997; Latif and Grötzner 2000) and Indian Oceans (e.g., Venzke et al. 2000). Predicting ENSO is therefore of considerable practical importance.

Recent studies have also shown that variability in the Atlantic and Indian Oceans can impact Pacific variability, and this knowledge could enhance ENSO prediction. In particular, sea surface temperature (SST) variations in these two regions may impact ENSO through atmospheric teleconnections (Dommengen et al. 2006; Jansen et al. 2009; Rodriguez-Fonseca et al. 2009). Thus, there is a two-way relationship between the Pacific and the Atlantic. However, the Atlantic Zonal Mode dominates the variations in the Equatorial Atlantic, and there is less direct impact by ENSO (Chang et al. 2006). Keenlyside and Latif (2007) showed that the correlation between SST anomalies in the central Equatorial Pacific and the central Equatorial Atlantic is strongest when the Atlantic is leading the Pacific with 6 months, implying a possible feedback from the Atlantic to the Pacific. Wang (2006) indicate that there is an inter-basin feedback that enhances the relation between the basins. Jansen et al. (2009) found that a cold (warm) Atlantic is associated with a warming (cooling) in the eastern Equatorial Pacific and a deepening (shallowing) of the average Equatorial Pacific thermocline about 6 months later. Frauen and Dommengen (2012) coupled an atmospheric general circulation model with a simplified ENSO recharge oscillator ocean model and found that the initial conditions of the Tropical Atlantic has a strong impact on ENSO predictability. Keenlyside et al. (2013) further show that observed ENSO predictions initiated in February can be enhanced by accurate knowledge of Equatorial Atlantic SST. North Tropical Atlantic SST may also influence ENSO variability (Ham et al. 2013). These studies suggest that the variability in the Atlantic Ocean may be important for both increasing the understanding of ENSO variability and predicting ENSO.

Rodriguez-Fonseca et al. (2009) support that there is a connection between the Equatorial Atlantic and the Pacific, with Atlantic SST anomalies leading Pacific SST anomalies by 6 months, although they also indicate that the strength of this connection varies in time. Based on observations, they propose that warm (cold) SST anomalies in boreal summer associated with the Zonal Mode events drive a strengthening (weakening) of the equatorial Trade Winds over the Pacific through the Bjerknes positive feedback (Bjerknes 1969) leading to La Niña (El Niño) like SST anomalies. Ding et al. (2012) reproduced the relationship and the teleconnection mechanism suggested by Rodriguez-Fonseca et al. (2009) with a partially coupled general circulation model. Recent studies (e.g., Keenlyside and Latif 2007; Rodriguez-Fonseca et al. 2009) have indicated a strengthening of the relationship between the Atlantic Zonal Mode and ENSO since the 1970s, but causes for a change in this relationship are still unknown. Several possibilities have been proposed: stochastic changes (Ding et al. 2012), or changes in the background state of the Pacific and the Atlantic (Rodriguez-Fonseca et al. 2009). Here we investigate if changes induced by a weakening Atlantic meridional overturning circulation (AMOC) can impact this relationship.

Model studies and sediment cores have indicated that an increased freshwater flux to the Arctic Ocean associated with melting of ice sheets can affect the general circulation in the atmosphere and the ocean, such as the thermohaline circulation in the Atlantic Ocean and the associated AMOC. Records also indicate that these changes in the AMOC could have played an important role in historic climate change (Broecker et al. 1985; Ganopolski and Rahmstorf 2001). In addition, numerous climate modeling studies have shown that an increase of greenhouse gas concentrations in the atmosphere leads to an increase of freshwater fluxes from the Arctic to the Atlantic (Räisänen 2001). This may in turn affect global climate through for instance changes in the AMOC (e.g., Stouffer et al. 2006). However, there are uncertainties and large variations in the estimates of the overall effect of an increased freshwater flow to the Atlantic in past and future climate scenarios (Schmittner et al. 2005; Kuhlbrodt et al. 2007).

Several recent studies (Dong and Sutton 2002; Zhang and Delworth 2005; Dong and Sutton 2007; Sutton and Hodson 2007), have investigated how an increased freshwater flux to the Atlantic and associated changes in the thermohaline circulation and AMOC can induce changes in ENSO. These studies have explained the response in ENSO by, for instance, atmospheric bridges over Central America and westward propagating Rossby waves excited in the Atlantic. All these studies identify significant changes in the mean state of the Tropical Pacific, and indicate that these induce ENSO changes.

In the present study, we investigate how a weakening AMOC, resulting from an increase in freshwater runoff from the Arctic region into the North Atlantic Ocean, and the associated change in the Atlantic mean state, can change the relationship between the Tropical Atlantic and Pacific regions. Our hypothesis is that a weakening AMOC may strengthen the relationship between Atlantic and Pacific variability. The present study is based on previous studies (Otterå et al. 2004; Breiteig 2009) using a global coupled atmosphere–ocean–sea ice model, the Bergen climate model (BCM).
The paper is organized as follows. In Sect. 2 the model and the data are described. The results are presented in Sect. 3. In Sect. 4 the results and possible mechanisms connecting the Atlantic and the Pacific are discussed and summarized.

2 Data: model and experimental design

BCM (Furevik et al. 2003), a coupled atmosphere–ocean–sea ice model, consists of the atmospheric general circulation model ARPEGE/IFS (Deque et al. 1994) and the ocean model MICOM (Bleck et al. 1992). The atmosphere model is run with a linear T63 grid. It has a horizontal resolution of about 2.8 × 2.8 degrees for physical processes and has 31 vertical layers from the surface up to 10 hPa. The ocean model has a non-isopycnic surface mixed layer on top of 23 isopycnic layers, and has a horizontal resolution of 2.4 × 2.4 degrees, which increases to 0.8 degrees near the equator. The ocean model is coupled with a dynamic and thermodynamic sea ice model. The atmosphere and ocean components are coupled with the OASIS coupler version 2.2 (Terray et al. 1995, 1998), and exchange data once every day. The heat and freshwater fluxes are adjusted based on a flux-correction to avoid drifting from the climatological SST and sea surface salinity fields (Furevik et al. 2003).

Since the BCM version used in this study is flux-adjusted, it is able to model an Atlantic cold tongue, something many general circulation models (GCM) fail to reproduce (Davey et al. 2002; Richter and Xie 2008). Hence, in BCM the Tropical Atlantic mean state and features of the Atlantic Zonal Mode are fairly well simulated compared to in many other GCMs (Fig. 1), as may be expected because the feedbacks responsible for the zonal mode depends on the mean state (Keenlyside and Latif 2007; Ding et al. 2010). However, Latif et al. (2001) showed that flux-corrected GCMs do not necessarily improve the state of the Tropical Pacific, but un-corrected models generally still simulate a cold tongue. In general, BCM simulates realistic ENSO characteristics, in strength, frequency and spatial pattern, but as common in many GCMs the Atlantic cold tongue is slightly too narrow and stretches too far west compared to observations (Furevik et al. 2003).

Otterå et al. (2004) performed a sensitivity experiment with BCM, consisting of two model simulations. One control simulation was carried out with greenhouse gas concentrations and aerosol particle concentrations kept constant at present-day values. This gave a simulated freshwater runoff to the North Atlantic Ocean of 0.1 Sv (1 Sv = 10^6 m^3 s^-1) equal to estimates of today’s observed climatological situation. A second so-called freshwater experiment (FW) was carried out with an artificially increased freshwater runoff of 0.4 Sv (Otterå et al. 2003, 2004). This increased runoff is of a size comparable to the runoff found for earlier deglaciations (Simonsen 1996) and also to model estimates for a situation where the greenhouse gas concentration of the preindustrial atmosphere is quadrupled (Manabe and Stouffer 1994). The control run was integrated for a 300-year period. FW was initiated at year 100 of the control run and integrated for 150 years. The freshwater flux increase in FW was located at coastal regions in the Nordic seas and the Arctic Ocean, and was added continuously throughout the 150-year integration. The freshwater flux was added artificially to the internal hydrological cycle, so the experiment can only be considered as a sensitivity experiment (Otterå et al. 2004).

In FW the AMOC weakens by about 6 Sv in the first third of the integration (Otterå et al. 2003, 2004). The strength of the AMOC, defined as the maximum overturning across the 24°N latitude belt in the Atlantic, for the control run (thick line) and FW (thin line) is illustrated in Fig. 2, where the grey box indicates the weakening phase. A weakening of the AMOC due to increased freshwater runoff to the Atlantic is consistent with findings from other GCM studies (e.g., Manabe and Stouffer 1997; Rind et al. 2001). However, BCM shows a weakening of the AMOC only in the first third of the integration period after which the AMOC gradually recovers to the same level as in the control run, whereas most studies show a reduction or a shutdown of the AMOC through the whole integration period (Otterå et al. 2004). Furthermore, the magnitude of the weakening is smaller in BCM compared to other similar studies, possibly related to a weaker freshwater forcing or different locations of the increased freshwater fluxes. Otterå et al. (2004) suggest that the weakened AMOC is caused by a reduction of the rates of deepwater formation in the Nordic Seas and the North Atlantic subpolar gyre, and a reduction of the southward flow of intermediate water masses through the Fram Strait. The strengthening of the AMOC with time is suggested to be caused by increased basin-wide upwelling in the Atlantic Ocean, northwards transport of saline waters from the western tropical North Atlantic, and a surface wind pattern maintaining the Atlantic water inflow between the Faroes and Scotland to the Nordic Seas (Otterå et al. 2004).

In this study we compare monthly mean data from the subperiod of FW where the AMOC weakens with a 150-year period of the control run that overlaps the period when FW was integrated. This 150-year subperiod in the control run is hereinafter referred to as CTRL. The sub-period in FW when the AMOC weakens is defined as the period from year 7 to 40 in FW. This 34-year period is referred to as DECLINE, illustrated by the grey box in Fig. 2. This DECLINE period is chosen because at year 7
the AMOC has achieved a relatively steep weakening rate, and by year 40 the AMOC shows the first tendency to strengthen again (Breiteig 2009). The strengthening AMOC phase is defined as the period from year 50 throughout the integration in FW, since this is after the AMOC reaches its minimum value in the maximum overturning. This recovering phase in FW is referred to as RCVR.

3 Results

3.1 Strengthened relation between Equatorial Atlantic and Pacific variability

To investigate the connection between the Equatorial Atlantic and Pacific, the cross-correlation between SST anomalies in the Atl3 region (3°S–3°N; 20°W–0°E) and the Niño 3.4 region (5°S–5°N; 170°W–120 W) is calculated for CTRL and DECLINE (Fig. 3). We define anomalies by removing the seasonal cycle. For comparison the corresponding cross-correlations for SST data from HADISST (Rayner et al. 2003) from 1870 to 2011 and 1970 to 2011, and for RCVR are included. For DECLINE correlations above 0.15 are statistically significant at a 95 % confidence level using a Fisher’s Z test.

For all datasets, there is a negative correlation when Atlantic SSTs lead Pacific SSTs with about six months, and a positive correlation when Pacific SSTs lead Atlantic SSTs with about six months. The negative correlation, when the Atlantic leads the Pacific, is three times as strong in DECLINE compared to in CTRL. Using a Fisher’s Z-transformation, negative correlations stronger than -0.14 at a six-month lead of the Atlantic in DECLINE is significantly different from CTRL at a 95 % confidence level. To quantify the uncertainty, we calculated the cross-correlation between Equatorial Atlantic and Pacific SST anomalies for 100 randomly selected 34-year periods from CTRL. A 95 % confidence interval for the CTRL ensemble correlation for a six-month lead of the Atlantic is then [-0.15, -0.12]. We repeated the cross correlation for 1000 randomly selected 34-year periods from CTRL, and found that the 2.5-percentile (-0.22) is weaker than the correlation in DECLINE. We conclude that the correlation of -0.29 found in DECLINE is significantly stronger than what is found in CTRL. In contrast, the difference in correlation between DECLINE and CTRL is smaller when the Pacific is leading the Atlantic. The relation between Atlantic and Pacific SST anomalies in DECLINE is similar to that observed during the period 1970 till present. However, stronger correlations are observed when the Pacific leads. We suggest that there exists a physical explanation for the difference in the correlation between CTRL and DECLINE, rather than mainly being a statistical property.
3.2 Pacific and Atlantic mean state

In DECLINE there is a distinct shift in the frequency of ENSO (Breiteig 2009), to a more pronounced 2-year period (Fig. 4). The power spectra of SST anomalies in the Niño 3.4 region indicate a peak in frequency at about 36 months in CTRL, while in DECLINE the peak is shifted to 25 months. To test the strength of the biennial mode of ENSO in DECLINE, a one-year lag correlation of the yearly winter (December-January–February) SST anomalies in the Niño 3.4 region is calculated. The correlation coefficient of $R = -0.55$ is statistically significant at a 95 % confidence level ($p = 0.009$). The corresponding correlation coefficient for CTRL is $R = -0.06$. Monte Carlo simulations indicate that there is a likelihood of less than 0.06 % for a correlation coefficient as large as in DECLINE to be achieved by chance. We conclude that the shift in frequency is statistically significant. In addition, the ENSO frequency shifts back to the value in CTRL when the AMOC strengthens again (Breiteig 2009). However, the significance of these changes is difficult to assess given the shortness of our record (Wittenberg 2009). In the Tropical Atlantic there is seemingly no significant change in the power spectrum of the Atl3 index as a response to the weakening AMOC (not shown).

There is no significant difference in the mean state of the Equatorial Pacific between CTRL and DECLINE (Fig. 5). The correlation of the annual average SST, precipitation, zonal wind stress and thermocline depth in the Equatorial Pacific between CTRL and DECLINE is high for all variables, with a pattern correlation coefficient $R$ close to unity, and a small root mean square error. This illustrates that the shift in ENSO frequency may not be attributed to a change in the mean state of the Equatorial Pacific.

Figure 6 shows the standard deviation of monthly SST in the Atl3 region for CTRL, DECLINE and RCVR, where the long-term mean has been removed prior to the calculation. SST data from HADISST from 1870 to 2011 are also included. The seasonal cycle of Equatorial Atlantic SST variability in BCM is substantially stronger than observed. In DECLINE the variability has increased during boreal spring and summer compared to in CTRL. For all three BCM datasets there is a peak in variability in April that is not observed. However, in DECLINE and RCVR there is a second peak in boreal summer that tends to coincide with observations. The additional large June peak in DECLINE is not present in CTRL or RCVR, and the variability in June is 20 % larger in DECLINE than in CTRL. This increase in variability is statistically significant at a 90 % confidence level by a two-tailed F-test.
Furthermore, by taking the standard deviation of SST for June for 100 randomly selected 34-year periods from CTRL, we find that the standard deviation in DECLINE is larger than the 95th-percentile. We conclude that the increased variability in DECLINE is significantly larger than in CTRL at a 90% confidence level.

In DECLINE, the peaks in SST variability in April and June coincide with the months when the monthly mean Equatorial Atlantic SST response to the weakening AMOC, defined as DECLINE-CTRL, is at its maximum (shown for June in Fig. 7). The response shows that SSTs are higher in the central Atlantic region in DECLINE compared to in CTRL. There is also a slightly negative response in the eastern Equatorial Atlantic from April to September (not shown). Combined this indicates a stronger zonal SST gradient in the Equatorial Atlantic. Figure 7 also shows the June response for precipitation, zonal wind stress and thermocline depth. Thermocline depth ($z_{20}$) is defined as the depth of the 20 °C isotherm. There is a deepening of the thermocline in the central Equatorial Atlantic indicating an increase in ocean heat content. There is also a dynamically consistent increase in precipitation and wind stress in this region, coinciding with the location of the positive SST response. The warming in the upper Tropical Atlantic may be due to a slow-down of the North Atlantic subpolar gyre caused by the increased freshwater flux to the Arctic Ocean, suppressing deepwater formation. This slow-down increases the travel time in the Guyana Current, increasing the temperature and salinity in the upper 600 m of the water column here (Ottera˚ et al. 2003, 2004).

### 3.3 Connecting the Tropical Atlantic and Pacific regions

How does the atmospheric circulation communicate the response from the Atlantic to the Tropical Pacific, and which dynamical processes can explain the correlation between Atlantic and Pacific SST anomalies? We investigate the previously proposed hypothesis that boreal summer Atlantic SST can influence the Trade Winds over the Pacific, and that the Bjerknes positive feedback subsequently enhances these anomalies (Rodriguez-Fonseca et al. 2009; Ding et al. 2012).

A regression analysis is performed for SST and precipitation anomalies (Fig. 8), and surface wind and thermocline depth anomalies (Fig. 9) regressed on to the summer (June-July–August) Atl3 index. This is done for boreal summer (June-July–August) and the following winter season (December–January–February) for DECLINE and CTRL. Only the regions where the correlation is statistically significant at the 90% confidence level under a $t$ test are shown. The patterns in the Equatorial Atlantic in boreal summer are in good agreement with the observed Zonal Mode structure in both experiments (e.g., Ding et al. 2012).
For DECLINE, there is a negative relationship between the Atl3 summer index and SST, precipitation, zonal surface wind and thermocline depth in the Equatorial Pacific in boreal summer. Thus, in DECLINE when the Equatorial Atlantic is warm in summer, there is a tendency for enhanced Trade Winds over the western Pacific, lifting the thermocline to the east and decreasing SSTs. In DECLINE, in the following winter, the negative relationship between Equatorial Atlantic SST anomalies in summer and Tropical Pacific SST, precipitation, thermocline depth and surface wind strengthens. This is consistent with ENSO dynamics and the Bjerknes feedback. There is a negative relationship between the summer Atl3 index and central Tropical Pacific SST anomalies, west-to-central Tropical Pacific convective precipitation anomalies, east Pacific thermocline depth anomalies and surface wind in winter. In CTRL the relationship between the summer Atl3 index and the Equatorial Pacific is weak and shows no significant changes.
from summer to winter. For comparison, a similar regression analysis done on observational data is shown in Rodriguez-Fonseca et al. (2009).

4 Summary and discussion

Our results indicate that the change in the Atlantic due to a weakening AMOC can induce a stronger connection between the Atlantic and Pacific SSTs, with the Atlantic leading the Pacific by about half a year. In the Atlantic there is a peak in SST variability in boreal summer, while in the Pacific this peak is in boreal winter. In the model, when there is a cold (warm) anomaly in summer in the Atlantic, there is a tendency for a warm (cold) anomaly in the Pacific the following winter. The correlation when the Atlantic leads the Pacific is stronger in DECLINE than in CTRL. Thus, the increased variability in the Equatorial Atlantic during summer strengthens the Atlantic–Pacific connection.

Previous studies suggest warm Atlantic Zonal Mode events in summer strengthen the zonal atmospheric circulation, with rising air in the eastern Equatorial Atlantic and sinking air in the central Equatorial Pacific. The divergence shallows the equatorial thermocline and cools the SST in the eastern Pacific. These anomalies are amplified by a coupled Bjerknes feedback process, causing weak La Niña-like anomalies in the winter. Similarly, a cold-event in the Atlantic in summer induces warm anomalies in the Pacific the following winter (Rodriguez-Fonseca et al. 2009; Ding et al. 2012). In our case, the zonal SST gradient in the Equatorial Atlantic has increased in DECLINE. This response can be associated with a westward shift of the Walker circulation and a strengthening of the ascending branch over the western Equatorial Atlantic. Our results are consistent with the explanation for the connection between equatorial Atlantic and Pacific SST anomalies discussed in Rodriguez-Fonseca et al. (2009).

Rodriguez-Fonseca et al. (2009) investigated the connection between the Tropical Atlantic and Pacific regions in observations for the period before and after 1979, and found that the connection has increased in the later period compared to the former. In observations from before 1979 the Pacific ENSO seems to have no apparent relationship with the Atlantic Zonal Mode, similar to what is observed in CTRL. However, after 1979, the Atlantic and the Pacific seem to have opposite phases with the Atlantic leading the Pacific, as in DECLINE. Tokinaga and Xie (2011) showed that since the 1950s there has been a reduction in the zonal SST gradient in the Equatorial Atlantic and a decrease in
the interannual variability of the longitudinal SST gradient (i.e. a weakening of the Atlantic Zonal Mode and a weakening of the equatorial cold tongue). In this same period SSTs have increased over the whole Equatorial Atlantic. The weakening of the Zonal Mode can be explained by the associated deepening thermocline in the east and thus, reducing the thermocline feedback (Tokinaga and Xie 2011). This is opposite to what was simulated in DECLINE, where the SST gradient strengthened, the thermocline shoaled in the east, and the Zonal Mode variability increased. In DECLINE both the connection between the Equatorial Atlantic and Pacific has strengthened and the Equatorial Atlantic variability has increased, while in observations after 1979 the connection has also strengthened but the Atlantic variability has reduced. Thus, the mechanism found here likely does not explain the stronger Atlantic-Pacific relation observed since 1979.

The effect of changes in AMOC on the Tropical Atlantic is still debated (Wen et al. 2010). Even though our results are not consistent with observations, the mechanism seen in this study may be of importance in other time periods. Whether AMOC induced changes could contribute to the strength of the Equatorial Atlantic-Pacific relation at other times remains unclear due to the lack of observations, and more work is required to understand the impact of future AMOC changes. Our results still suggest that the Atlantic-Pacific relationship depends on the state of the Tropical Atlantic, and that seasonal ENSO forecasts and the understanding of ENSO variability can be improved by including Tropical Atlantic surface conditions (Keenlyside et al. 2013).

In DECLINE, when the relation between Atlantic and Pacific SSTs is strengthened, the power of the two-year period of ENSO is also increased (Fig. 4). Thus, the strengthened variability in the Equatorial Atlantic during summer could potentially enhance an intrinsic dominant ENSO mode of the Pacific, strengthening the two-year ENSO period in DECLINE. Similar freshwater hosing experiments to the one presented here using other GCMs, have indicated an effect of Atlantic SSTs on ENSO (Dong and Sutton 2002; Timmermann et al. 2005; Zhang and Delworth 2005; Dong and Sutton 2007; Sutton and Hodson 2007), but in contrast to these studies, our results show a response in ENSO that may not be attributed to a response in the Tropical Pacific mean state. The fact that we do not find significant responses in the Tropical Pacific mean state may be related to the weaker freshwater forcing and the resulting weaker response in the AMOC in our study compared to other similar freshwater hosing experiments. In addition, several of these previous studies have not commented directly on the connection between Equatorial Atlantic and Pacific SST anomalies, and the connection may not be present in these studies. The reason for this may be due to the poor simulation of Equatorial Atlantic variability in many present GCMs. As previously mentioned, the BCM version used in this study is flux-adjusted, and the Tropical Atlantic is therefore fairly well simulated compared to in many other GCMs (Davey et al. 2002; Richter and Xie 2008). However, it should be noted that flux-adjustments might change model sensitivity to freshwater perturbations, although Schiller et al. (1997) states that this may not be of significance.

Although we have not explored changes in the structure of ENSO events in this paper, there are indications that ENSO shifts to a more central-Pacific (CP) type ENSO in DECLINE compared to a more eastern-Pacific (EP) type ENSO in CTRL. The CP-ENSO has been associated with shorter oscillation periods (Kao and Yu 2009) and SST anomalies displaced to the west in the Tropical Pacific compared to the EP-ENSO (Kao and Yu 2009; Yeh et al. 2009). A shift from an EP-ENSO to a more CP-ENSO could explain the difference in ENSO frequency between DECLINE and CTRL. We also find SST anomalies shifted farther west in DECLINE compared to in CTRL consistent with the EP-CP definitions (not shown). Previous studies (e.g., Ham et al. 2013) have argued that North Tropical Atlantic SST anomalies in spring can influence the frequency of CP events. Our model results suggest that Equatorial Atlantic SST in summer might also be important in this respect. The mechanism for how the Equatorial Atlantic variability can change the frequency of CP events is however unclear. We hypothesize that external forcing from the Atlantic may enhance an innate ENSO-mode with a two-year period.

To summarize, in BCM the relationship between the Equatorial Atlantic and Pacific depends on the mean state of the Tropical Atlantic. This study has shown that when the AMOC weakens (1) the variability increases in the Atlantic in boreal summer, possibly due to a stronger zonal mean SST gradient, (2) Atlantic Zonal Mode events tend to lead opposite signed ENSO-like anomalies by about six months, and (3) the two-year oscillation period in the Equatorial Pacific is more pronounced, while the mean state of the Pacific does not show significant changes. Regression analysis showed that Equatorial Atlantic and Tropical Pacific SST, surface wind and thermocline depth anomalies are related in boreal summer, and these anomalies are subsequently amplified in the Pacific by the Bjerknes positive feedback, accounting for the six-month lead of the Atlantic. Under control conditions the relation between Equatorial Atlantic and Pacific boreal summer anomalies is insignificant. Thus, the weakening AMOC triggers a stronger coupling between the Tropical Atlantic and Pacific regions.

We acknowledge that the length of DECLINE is quite limited, and that interannual variability in the Equatorial
Pacific may not have statistical robustness in short time series. The existence of a physical mechanism provides some confidence in our results. However, an ensemble of AMOC weakening experiments is required to assess the robustness of our findings. Nevertheless, we hope that our results motivate further work in this direction with more models.

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Paper V

Investigating the role of the Atlantic and Pacific in the early 20th century warming

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Manuscript in preparation
Investigating the role of the Atlantic and Pacific in the early 20th century warming

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Abstract

Instrumental records show that there have been two periods of enhanced Northern Hemisphere warming in the 20th century, the early warming from 1910-1940 and a later period from around 1970. State-of-the-art coupled models have difficulties simulating the early 20th century warming, and the causes of these variations are not well understood. To improve our understanding of the early 20th century warming, we have performed several ensemble experiments with a global coupled model, the CMIP5 version of the Norwegian Earth System Model (NorESM). The experiments consists of ensembles of historical all-forcing 20th century simulations where daily momentum flux anomalies are prescribed to the ocean globally, or only to the Atlantic or the Indo-Pacific region. Using this method we are able to phase the sea surface temperature variability to observed, while keeping the system thermodynamically coupled, and since all simulations include the same historical forcing we can to some extent separate the externally forced signals from dynamically driven variability. We find that the early 20th century warming in the Northern Hemisphere is only simulated when momentum flux is prescribed in the Indo-Pacific, indicating that the phasing of decadal variability of the Atlantic is not necessary for simulating this warming, with decadal variability in the Pacific having a larger contribution.

1. Introduction

The global mean surface temperature has been increasing for the last century. Multi-decadal variability is superimposed on to this trend with periods of enhanced warming. However it is still unclear what drives this variability on decadal timescales. During
the 20th century there were two periods of increased warming, the early 20th century warming between 1910 and 1940, and the recent warming beginning around year 1970. While the latest warming period has been attributed to anthropogenic influence (e.g., Tett et al. 1999; 2000; Crowley 2000; Stott et al. 2000; Meehl et al. 2004), the causes of the early warming are still not fully understood. Here we investigate the roles of variability in the different ocean basins in the early 20th century warming in the Northern Hemisphere, including the Arctic.

Earlier studies have proposed a number of possible explanations for the early 20th century warming, ranging from atmospheric greenhouse gas (GHG) concentrations and natural climate forcing to combinations of these, as well as contributions from natural internal variability (e.g., Tett et al. 1999; 2002; Delworth and Knudson 2000; Stott et al. 2000; Broccoli et al. 2003; Hegerl et al. 2003; Meehl et al. 2004). When it comes to internal variability, a discussion is ongoing about the importance of the Atlantic (e.g., Schlesinger and Ramankutty 1994; Zhang et al. 2007; Steinman et al., 2015) versus the Pacific for this warming (e.g., Dai et al. 2015; Thompson et al., 2015). During the early 20th century warming period there was low volcanic activity and intensified solar radiation (e.g., Crowley 2000). Atmospheric GHG concentrations were increasing, but not at the rate of today. However the 20th century warming might have been too strong to be driven only by these external factors, and internal variability possibly played a role (Keenlyside and Ba 2010). For instance during this same period there is evidence of a positive phase of both the Pacific Decadal Oscillation (PDO) and Atlantic Multi-decadal Variability (AMV) although the warming in the Atlantic was delayed by a decade compared to the global temperature increase. A similar discussion debating the roles of external forcing and internal variability, especially the phasing of decadal variability in the Pacific, is ongoing for the recent so-called Hiatus, where the positive global mean surface temperature trend has flattened (e.g., Salomon et al. 2010; Koseka and Xie 2013; Trenberth and Fusallo 2013; Fyfe and Gillett 2014; England et al. 2014).
Delworth and Knudson (2000) investigated global temperature variability in a six-member ensemble of a fully coupled global climate model with time varying historical forcing, and found that only one of the ensemble members simulated the observed early 20th century warming. They concluded that the early warming was due to a combination of internal variability and anthropogenic radiative forcing. Steinmann et al. (2015) analyzed a large amount of model realizations from the Coupled Model Intercomparison Project Phase 5 (CMIP5; Taylor et al. 2012) and found that AMV and PDO together explain a large part of the internal temperature variability over the Northern Hemisphere. They suggest that a combination of a strong cooling in the Pacific combined with only a weak warming in the Atlantic has lead to the recent slowdown of global warming, while internal variability in the Atlantic played a more important role in the early and middle 20th century. Similarly, Zhang et al. (2007) found that by only constraining North Atlantic temperature fluctuations to the observed they were able to reproduce the observed multi-decadal variability in Northern Hemisphere surface temperatures. In contrast, Dai et al. (2015), also using a large number of CMIP5 models, found that internal variability mainly associated with the PDO has been the main driver of accelerations and decelerations of global temperature trends. Thompson et al. (2015) support this finding, through enhancing the data availability in the tropical Pacific by reconstructing wind and sea surface temperatures (SSTs) from corals in this region. They find that tropical Pacific trade winds were weaker and SSTs were higher during the early 20th century warming period, and that the situation was reversed during the latter period from 1940-1970 (Thompson et al. 2015).

There are also indications that the Arctic warmed substantially during the early 20th century warming period. Bengtsson et al. (2004) suggested that the Arctic warmed due to warm water advection into the Barents Sea (Bengtsson et al., 2004), indicating the importance of the Atlantic inflow. However, Suo et al. (2013) found that low volcanic activity during the period of the early 20th century warming in combination with increased solar radiation were more important for the warming in the Arctic. The Pacific has also been suggested as a possible source for the Arctic early warming, with
the PDO transitioning from a warm phase to a cold phase in the 1940s (Yamanouchi 2011).

To this extent it is clear that a number of different factors could be important for the early 20th century warming. Models differ in their climate sensitivity to for instance solar and GHG forcing, relaxation time to volcanic eruptions, and magnitude of internal variability, so that the fractional contributions of each of these forcing factors vary between models. In addition, data are scarce during this early warming period, especially for the oceans (Smith and Reynolds 2003), so more model studies as well as proxy reconstructions are needed to clarify these differences. We hypothesize that the phasing of decadal variability in the ocean is an important driver of the early 20th century warming in the Northern Hemisphere and we investigate the roles that decadal variability in the Atlantic and Pacific played in this warming in a single model framework.

2. Data and Methods

To investigate the roles of the Atlantic and the Pacific in the early 20th century warming, we have performed several ensemble experiments with the Norwegian Earth System Model (NorESM). NorESM consists of the atmospheric component CAM4-Oslo, the ocean component MICOM, the land component CLM4, the sea ice model CICE4, and the coupler CPL7. NorESM is based on the Community Earth System Model (CESM1; Lindsay et al. 2014), with the same land and sea ice components and coupler, but differs in the following aspects. NorESM uses an isopycnic coordinate ocean general circulation model developed from MICOM (Bleck et al. 1992), and the atmosphere component includes a different chemistry-aerosol-cloud-radiation interaction scheme. The NorESM1-ME version used here also includes prognostic biochemical cycling, which is deactivated in this study. Here we use the CMIP5 version of the model. See Bentsen et al. (2012) and Tjiputra et al. (2013) for more details.
We have performed four six-member ensembles with NorESM with transient 20th century historical forcing (Table 1). First we have a control ensemble (CNTRL) with six members of fully coupled historical simulations as done for the CMIP5. We initialize the historical simulations at year 1850 with initial conditions given by a preindustrial control simulation. The initial conditions are selected at a 10-year interval from the preindustrial control simulation because internal variability in the Atlantic in NorESM has a periodicity of about 20 years (Bentsen et al. 2012).

The other three ensemble simulations are partially coupled, similar to the setup from Ding et al. (2014a). We prescribe daily momentum flux anomalies from the 20th Century Reanalysis data provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their Web site at http://www.esrl.noaa.gov/psd/ (20CR; Compo et al. 2011) to the ocean component of the model. The 20CR anomalies are added to the models daily climatology, in contrast to a fully coupled simulation where the ocean component receives momentum flux from the atmosphere component through the coupler. The anomalies and climatology are both calculated based on the years 1901-2000, where the climatology is calculated from the CNTRL ensemble. Since we are prescribing anomalies from 20CR the mean state of the model does not change. This method is used to constrain the model variability to observations, while at the same time maintaining an active thermodynamic coupling. In this way SST is a fully prognostic variable that can freely interact with the atmosphere component of the model. Since all four ensembles include the same transient forcing, we can use this experimental setup to separate the dynamical driven ocean changes from the radiative forced changes.

In the second ensemble (TAU) we prescribe the momentum flux globally at every grid point over the ocean. In grid boxes where there is sea ice we weight the momentum flux with the ice fraction, so that for grid boxes with 100% sea ice the model is fully coupled. We initialize these simulations at year 1900 from the corresponding ensemble member in CNTRL, and the momentum flux modifications are implemented from year 1901. Thereafter we run the model for 100 years, giving us an ensemble spanning
years 1901 to 2000. In addition we have performed two partially partial coupled ensemble simulations similar to TAU with the same initial conditions, where we have only prescribed momentum flux in the Atlantic (TAU-ATL) or the Indo-Pacific (TAU-PAC), respectively. Otherwise the model is fully coupled. In TAU-ATL we prescribe momentum flux from 25°S to 65°N in the Atlantic and in TAU-PAC from 25°S to 60°N in the Indo-Pacific. We have a tapering region of linear weighting of 5 degrees latitude outside these boundaries. For the Atlantic the tapering regions are between 35°S and 30°S at the southern boundary and 65 and 70°N at the northern boundary to include the sub-polar gyre, but exclude the Barents Sea from the region of prescribed momentum flux. For the Indo-Pacific the tapering region at the northern boundary is from 60 to 65°N at the Bering Strait, while the southern region is the same as in TAU-ATL, at the southern tip of Africa. The area that is partially coupled is shown in black in Figure 1, with the grey areas being the tapering zones. White areas are where the model is fully coupled.

We chose the momentum flux from the 20CR because it was the longest available product at the time these experiments were initiated. Furthermore, we have chosen to use the ensemble mean momentum fluxes, anticipating them to be the most accurate. However, the ensemble mean is known to have less variance during the earlier period when observational data are scarce (He et al. 2016); the implications of this choice are discussed below. In future work, we plan to check the sensitivity of the results to this and also other available long-term reanalysis (e.g., ERA20C). We also use the Hadley Centre sea ice and sea surface temperature data set (HadISST; Rayner et al. 2003), the gridded sea level pressure dataset HadSLP2 (Allan and Ansell 2006), the Goddard Institute for Space Studies Surface temperature analysis (GISTEMP; Hansen et al. 2010) and other atmospheric variables from the 20CR to compare with our model results.

These experiments are here used to investigate the impact of the Atlantic and Indo-Pacific on the early 20th century warming period from 1910-1940 over the Northern Hemisphere including the Arctic.
3. Results

3.1 Validation of our experimental setup for investigating the early 20th century warming

The aim of our model set-up is to constrain the model to observations, while maintaining the thermodynamic coupling. In this way we can constrain the phasing of the climate variability over the oceans. The correlation patterns between simulated and instrumental de-trended annual mean SST show that to a large degree we are successful on interannual timescales (Figure 2). In CNTRL the correlation of de-trended data is low globally as simulated internal variability in the model is not in-phase with observations. For TAU we have significant correlations globally, especially high in the tropical Pacific, by simulating the observed timing and variability of El Niño-Southern Oscillation (ENSO) events (Figure S1). The correlation patterns for TAU-ATL and TAU-PAC match the masking of the basins, with the Indo-pacific region simulating observed SST variability in TAU-PAC, and the North Atlantic simulating observed SST variability in TAU-ATL. In TAU-PAC there is also an area with significant correlation in the tropical Atlantic, similar to the ENSO teleconnection pattern in this region (e.g., Alexander et al., 2002). However we are not able to constrain the South Atlantic, possibly due to the large model biases that exist in this region (Richter and Xie, 2008, Ding et al., 2015). The poor simulation of interannual variability and also decadal variations over large parts of the Southern Hemisphere (Figure S2) could in part be due to uncertainties in the observation products, particularly prior to the 1950s (He et al. 2016). Therefore our focus here will be on the trends in the Northern Hemisphere.

It is a challenge for present day climate models to simulate the early 20th century warming (e.g., Delworth and Knudson 2000), and CNTRL simulates an increase of less than 0.2 K in Northern Hemisphere surface temperature compared to the 0.5 K increase in the GISTEMP data. Figure 3 shows the 11-year running mean of area-weighted averaged Northern Hemisphere surface temperatures for the four ensembles and for the GISTEMP data. While all four ensembles simulate the observed later warming starting around year 1970 alluding to the importance of external forcing
during this period, only TAU and TAU-PAC, where we include momentum flux
anomaly forcing to constrain the variability in the Pacific, are able to simulate a
warming comparable to the GISTEMP data, of about 0.4 K, and the observed warm
peak in the Northern Hemisphere around 1940, significantly warmer than in CNTRL.
A less pronounced early warming period can be discerned in TAU-ATL with a total
warming of 0.3 K. The Northern Hemisphere surface temperature in TAU-ATL peaks
a decade before the observed data, while in TAU and TAU-PAC the peaks are a
couple of years delayed compared to the observed data. There are also discrepancies
between the simulations and the GISTEMP data in the first 10 years of the 20th century
that could be partly from the spin-up of the model to the imposed 20CR fluxes from
year 1901.

By constraining SST variability in the Pacific with momentum flux from the 20CR, we
are able to simulate multi-decadal Northern Hemisphere surface temperature
variability; especially we improve the simulation of the 20th century early warming,
but TAU and TAU-PAC overestimate the cooling trend from 1940-1960. The cooling
after the 1940s is more uniform across latitudes in TAU and TAU-PAC (Figure 4).
This exaggerated cooling trend results in a weaker overall Northern Hemisphere
warming in TAU and TAU-PAC for the total 20th century. The model simulations are
warmer than the GISTEMP data for the first half of the 20th century; however there
might be a cold bias in the observed data during this period (Karl et al. 2015).

3.2 Early 20th century warming in the Arctic
In GISTEMP the zonal mean surface warming is strongest in the Arctic, with a weaker
trend in the Northern Hemisphere midlatitudes and the tropics (Figure 4). A similar
pattern with warming in the Arctic is present in TAU, TAU-ATL and TAU-PAC. In
TAU and TAU-PAC the warm anomalies in the tropics are larger than in the
GISTEMP data, while in TAU-ATL the warming is confined to higher latitudes. In
CNTRL the warming is generally weak.
The surface temperatures in the Arctic in GISTEMP increase in phase with, but to a larger degree than the Northern Hemisphere temperature from 1910-1940. While the Northern Hemisphere temperatures increase with around 0.5 K, the Arctic warms with more than 1K. Figure 5 shows the 11-year filtered (running mean) annual mean area-weighted average surface temperature over the Arctic from 70-90°N for the four ensembles and the GISTEMP data. For the model data only grid points where there is coverage in the GISTEMP data are included. Including these grid points does not notably change the results from those presented here. All four of our ensembles simulate the later warming from 1970, however TAU appears closest to the observed changes, being significantly cooler than CNTRL. Prior to this period, CNTRL exhibits the weakest decadal trends, underestimating the early 20th century Arctic warming by up to 50%. TAU and TAU-PAC better manage to capture the trend in the early 20th century warming period in the Arctic, with temperatures increasing by about 1K for both TAU and TAU-PAC. In TAU-ATL the Arctic warms by less than 0.8K, and the anomalies are only significantly different from CNTRL for the first half of the early 20th century warming period. Both TAU and TAU-PAC are significantly warmer than CNTRL in the later part of the warming period.

Although data are scarce for the Arctic for the early 20th century warming period, the GISTEMP data show a warming trend around the Bering Strait and in the Barents-Kara Sea and especially over Greenland (Figure 6). None of the ensembles simulate such a strong warming over Greenland, but both TAU and TAU-ATL have a weaker but significant warming trend here. TAU and TAU-ATL also have a strong warming in the Barents-Kara Sea, significantly different from CNTRL. All three experiments have a warming in the eastern Arctic and over Siberia. The trends in sea ice cover over the Arctic match the trends in surface temperature with the strongest decrease in sea ice cover in the Barents-Kara Sea. The temperature trend in the Arctic is largest during the winter from December to February (DJF) in the GISTEMP data. This seasonality is not as pronounced in the experiments.
A separation of the Arctic into an Atlantic part (90W-90E) and a Pacific part (90E-90W), shows that even though both TAU and TAU-ATL simulate a strong warming trend in the Barents-Kara Sea, the experiments are not able to simulate the observed strong trend on the Atlantic side of the Arctic (Figure 7). However the observed trend is well captured in TAU and TAU-PAC on the Pacific side, increasing by 1K during this period. Here TAU-ATL and CNTRL are not distinguishable after year 1920, with a temperature increase of less than 0.5K for both simulations.

Since the warming trend in the Barents-Kara Sea is only simulated in TAU-ATL and TAU, this seems related to the variability in the Atlantic, while the trends on the Pacific side that are only reproduced in TAU-PAC and TAU are related to changes in the Pacific. In the following we will investigate further the variability in the Atlantic and Pacific to identify their roles in the Arctic warming trend and also for the whole Northern Hemisphere.

3.3 The role of decadal variability in the Atlantic and the Pacific

The observed high-latitude warming pattern for the period 1910-1940 (Figure 6a) coincides with a PDO-like (e.g., Mantua and Hare 2002) warming pattern over the North Pacific and a basin-wide warming over the North Atlantic, resembling the positive phase of AMV (Figure 8a). The pattern in CNTRL does not resemble the pattern in GISTEMP, and the warming is overall much weaker. The pronounced Northern Hemisphere warming from 1910-1940 simulated in TAU and TAU-PAC coincides with a PDO-like warming over the North Pacific similar to observations, but differences are found over land. All three experiments, TAU, TAU-PAC and TAU-ATL, have significant warming trends over Siberia. In TAU-ATL this seems to extend from the warming maximum in the Barents-Kara Sea, while in TAU and TAU-PAC the maximum is near the Bering Strait. The North Atlantic also warms significantly in TAU and TAU-ATL, and in TAU-ATL there is a warming trend over North America. The warming over the North Atlantic is stronger in TAU than in TAU-ATL possibly due to impacts from the Pacific. However all ensembles underestimate the warming
over Greenland. As expected, TAU-ATL simulates no PDO-like warming during this period, and TAU-PAC simulates no substantial warming over the North Atlantic.

The observed, reanalysis, and simulated 1910-1940 trends in annual sea level pressure (SLP) (Figure 9) and the 200 hPa geopotential height (Figure 10) show similarities to the trend in surface temperatures (Figure 8). Especially a positive trend in 200 hPa geopotential height and SLP over the northern North Pacific is apparent in TAU and TAU-PAC, with significant pattern correlations in the Northern Hemisphere between 200 hPa geopotential height and surface temperature of 0.66 in TAU and 0.58 in TAU-PAC. Similarly the pattern correlation between 200 hPa geopotential height and SLP is 0.79 in TAU and 0.65 in TAU-PAC. In TAU-ATL there is a wave train going eastward from the North Atlantic, and a negative trend in SLP over Europe bringing warm air northwards towards Siberia and the Arctic. In TAU and TAU-PAC there is a stronger positive trend in SLP in the North Pacific. This anti-cyclonic tendency indicates a weakening of the Aleutian Low, with a tendency for more warm air from over the North Pacific moving northwards in over Siberia and the Arctic. There is a similar trend over the North Pacific in the HadSLP2 data, although the signal is weaker.

Even though the Northern Hemisphere surface temperature variability is somewhat constrained in our experiments (Figure 2 and Figure 8), there is an inter-member spread in the trend pattern for surface temperature, SLP and upper level geopotential height over land (not shown). However for TAU all ensemble members simulate a warming over the North Atlantic and parts of Eurasia, as well as the North Pacific and south of the Bering Strait. In TAU-PAC all members have a positive PDO-like signal in the surface temperature trend and a warming over Siberia similar to GISTEMP and in TAU. Northeastern Eurasia and North America warms in all members of TAU-ATL as in GISTEMP. Two members of TAU-ATL also have warming trend patterns in the Pacific resembling a positive PDO signal, similar to the patterns in TAU, TAU-PAC and in the GISTEMP data. These two ensemble members have a stronger early 20th century warming in the Northern Hemisphere than the TAU-ATL ensemble mean.
In the observed data, both AMV and PDO change phase from negative to positive during the early warming period. A simple linear regression model for multi-decadal Arctic surface temperature shows us that using AMV, PDO and the linear Northern Hemisphere surface temperature trend line as predictors we can explain up to 80% of the variance in Arctic surface temperature, where AMV and PDO are about equally weighted.

The PDO-index from observed data and the ensembles are shown in Figure 11. We define the PDO as the low-frequency filtered first principle component of detrended SSTs between 20-60°N and 120°E-120°W (e.g., Mantua and Hare 2002). The observed PDO-index increased from 1910 until the 1940s, but in both TAU and TAU-PAC the PDO lag the observed data by 5-10 years in the first half of the century. The maximum correlations between the observed PDO and the simulated PDO in both TAU and TAU-PAC during the first 50 years of the simulations are when the observed leads by six years (r=0.88). The PDO in TAU and TAU-PAC follow the observed PDO well in the second half of the century, with the maximum correlation at lag zero (r>0.9), changing sign and peaking again in the early 1980s. The delayed PDO in the first half of the simulation could be due to the response time of the ocean to the prescribed momentum flux that starts in year 1901. In both CNTRL and TAU-ATL there is a spread in the PDO in the ensemble members likely due to internal variability, but for TAU and TAU-PAC the PDO-indices have similar phasing for all realizations, consistent with a wind-driven PDO (e.g., Schneider and Cornuelle 2005). The timing of the phase change of the PDO from negative to positive in TAU and TAU-PAC match the early warming period in these ensembles, with a slight delay compared to the observed data.

In phase with the Northern Hemisphere warming trend, the observed AMV transitions from a negative to a positive phase (Figure 12). Here we define the AMV-index as the 11-year filtered (running mean) annual average of North Atlantic SST anomalies between the equator and 60°N, and from 75°W to 7.5°W (Enfield et al. 2001; Wyatt et
al. 2012). The simulated AMV in TAU and TAU-ATL, and somewhat in TAU-PAC, also transition from a negative to a positive phase during the early 20th century warming, but we are not able to simulate the observed phasing of the AMV after the 1950s, and the amplitude of the modeled AMV is too small. The amplitude of AMV is about 0.2 K in the observed data, while it is half the size in the ensembles. The warming period of the AMV in TAU and TAU-ATL is also shorter than the Northern Hemisphere early warming period, warming from the 1920s and starting to cool already in the 1930s. The correlation between the observed AMV and the simulated ensemble AMV is 0.67 for TAU and 0.40 for TAU-ATL, while the correlations are insignificant for the other ensembles. Although earlier studies have suggested an atmospheric driven part of AMV (e.g., Eden and Jung 2001; Mecking et al. 2013), the prescribed 20CR momentum flux is not able to force the strength of the AMV to observed in NorESM, compared to a clearer wind-driven multi-decadal signal in the Pacific.

Since there are similarities in the phasing of AMV between TAU-PAC and TAU and TAU-ATL, while not with the AMV in CNTRL, the AMV seems to have a Pacific-forced component. This can also be seen in the correlation pattern for TAU-PAC in Figure 2, where there is a significant correlation in the tropical North Atlantic. Regressing the first principle component of the low-pass filtered curl of the prescribed wind stress anomalies onto global SSTs in TAU does give us a pattern that resembles a PDO and the tropical Pacific teleconnection pattern (Figure S3). This pattern is similar for TAU-PAC, while in TAU-ATL significant values are mostly confined to the Atlantic (not shown). Although not the focus of this study, other studies have found that the Atlantic and Pacific can interact on multi-decadal timescales (Latif 2001; Zhang and Delworth 2007).

4. Discussion

While all four ensembles simulate a centennial warming trend in the Northern Hemisphere, and an acceleration of this trend from around year 1970 as in instrumental records, only the ensembles with prescribed momentum flux simulate a
pronounced early 20th century warming. This indicates that external forcing alone is not the driver for the early 20th century warming, at least in NorESM.

As the Northern Hemisphere warms in the observed data, the PDO changes sign from a negative to a positive phase. The AMV also shifts from a negative to a positive phase during this period, but here the warming begins later than in the Pacific. In both TAU and TAU-PAC we are able to simulate the observed PDO, and the PDO is in phase with Northern Hemisphere and Arctic surface temperatures. There is a positive PDO-like pattern in the early warming trend over the Pacific in the GISTEMP data as well as in TAU and TAU-PAC. The atmosphere circulation in the North Pacific mirrors the surface temperature trend, with anticyclonic flow bringing warm extra-tropical air into the Arctic. Only in TAU and TAU-PAC are we able to also simulate the observed warming trends on the Pacific side of the Arctic, while the warming is too weak on the Atlantic side in all simulations. Li et al. (2015) found the opposite atmospheric pattern over the North Pacific for the cooling of the Bering Strait region of the Arctic between 1994-2013, with negative geopotential height anomalies at 300-100 hPa. TAU and TAU-ATL also simulate a significant warming in the North Atlantic during the early warming period. This warming coincides with a phase change of the AMV, as in observational data. Nevertheless we are only able to simulate the observed warming peak in Northern Hemisphere and Arctic surface temperatures when the Pacific is constrained and we simulate the observed phasing of the PDO, and we conclude that decadal variability in the Pacific contributes to the early warming trend. Supporting this conclusion, two ensemble members of TAU-ATL simulate a warming trend in the North Pacific in the early warming period and then also have a stronger early warming trend than the TAU-ATL ensemble mean (not shown).

Even though the trend pattern in the Pacific in TAU and TAU-PAC are similar to observed, the trend pattern for atmospheric circulation over land, especially over Eurasia, differ between ensemble members and compared to observations, indicating the role of internal climate variability here. A similar result is found in Kosaka and Xie (2013), where they are able to simulate the recent observed global temperature hiatus,
but the temperature patterns over Eurasia differs from observations. They attribute this to that temperature variability in this region is not related to forcing from the Pacific (Kosaka and Xie 2013). Also the strong warming observed over Greenland is only present in a few ensemble members, indicating that this could also be due to regional internal atmospheric variability following Ding et al. (2014b).

It is difficult to isolate the contribution from the Pacific in the early warming, since the system is partially coupled, and the Atlantic seems to have a Pacific-forced component. The more persistent warming in TAU and TAU-PAC compared to TAU-ATL does indicate though that the phasing of decadal variability in the Pacific is enough to force the early 20th century warming. The phasing of the AMV could contribute to the Northern Hemisphere warming as well, as the warming in TAU-ATL is stronger than in CNTRL. However the amplitude of the simulated AMV is too weak compared to observed data and the warm phase of the AMV does not persist long enough. We can speculate that if we had been able to simulate the AMV more realistically with a longer and warmer warm phase, the early warming in TAU-ATL could have been stronger. The simple linear regression model indicated both the AMV and PDO as predictors for decadal variability in the Northern Hemisphere. However, the Pacific and Atlantic contribute to the warming in a non-linear fashion, with the trend patterns in TAU not being a combination of the trend patterns in TAU-ATL and TAU-PAC. Nevertheless since the early warming in both TAU and TAU-PAC is similarly strong as in observations, the AMV does not seem necessary for the early 20th century warming trend.

Even though we are not able to simulate the full extent of the early warming in TAU-ATL, the warming in the Barents-Kara Sea during this period as seen in earlier studies (Bengtsson et al. 2004; Suo et al. 2013) is present in both TAU and TAU-ATL. This regional warming seems therefore to be related to changes in the Atlantic, for instance through Atlantic inflow, with decreasing sea ice here and heat content in the North Atlantic increasing during this period. Earlier studies have suggested both external forcing and Atlantic inflow as drivers for warming in the Arctic during the early
warming period (e.g., Bengtsson et al. 2004, Suo et al. 2013). Since this regional warming is not present in CNTRL, we conclude here that it is likely due to Atlantic inflow and not externally forced.

In NorESM external forcing alone is not able to simulate the early 20th century warming, however some models are more sensitive to external forcing, with the decadal variability being phased by this (Otterå et al. 2010, Booth et al. 2012). Suo et al. (2013) investigated the early 20th century warming in the Arctic in a different model, the Bergen Climate model, and concluded that external forcing was an important factor for driving this warming. Otterå et al. (2010) using the same model to investigate the drivers of AMV concluded that external forcing is also important for the phasing of the AMV. We speculate that the early warming on the Atlantic side of the Arctic, at least in the Barents-Kara Sea, is in part related to AMV, either it is external forced or internal variability.

The global mean surface temperature has in the recent 15 years experienced a so-called Hiatus, where the positive global mean surface temperature trend has flattened. Although some studies have suggested that this hiatus is forced by a combination of external factors, such as atmospheric water vapor and aerosol content in combination with a solar minimum (Fyfe and Gillett 2014; Schmidt et al. 2014; Solomon et al. 2010), recent studies have suggested that decadal variability in the Pacific is the major driver, cooling over the last decade (Kosaka and Xie 2013; Trenberth and Fusallo 2013; England et al. 2014; Steinmann et al. 2015). Similarly we find that decadal variability in the Pacific could have contributed to the early 20th century warming in the North Hemisphere and in the Arctic.

As already mentioned, these results are model dependent. NorESM could be less sensitive to external forcing than other models, and the Pacific teleconnection pattern might be too strong. The observational data for the early warming period are also scarce making comparisons with observations limited as well. Therefore similar
experiments will have to be performed with other models forced by other reanalysis products.

5. Conclusions
In this study we have used a novel approach for separating the radiative forced and the dynamically driven components in a coupled climate system to identify drivers of the observed early 20th century warming in the Northern Hemisphere including the Arctic. By partially coupling transient 20th century historical forcing simulations of the NorESM with prescribed momentum flux anomalies from 20CR over the ocean, we find that half of the warming observed in the early 20th century is dynamically forced through decadal variability in the Pacific, in addition to the radiative forced part. A part of the dynamically driven forcing could potentially also be attributed to the Atlantic, however since we can reproduce the early 20th century warming without a pronounced AMV signal, we conclude that the phasing of decadal variability in the Atlantic is not key to the early 20th century warming in the Northern Hemisphere and the Arctic.

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References


Table 1 Description of ensemble simulations. All ensembles have 6 members each and are forced with the 20\textsuperscript{th} century historical transient forcing as in the CMIP5.

<table>
<thead>
<tr>
<th>Ensemble Experiment setup</th>
<th>CNTRL</th>
<th>TAU</th>
<th>TAU-ATL</th>
<th>TAU-PAC</th>
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<tbody>
<tr>
<td></td>
<td>Fully coupled</td>
<td>Partially coupled globally</td>
<td>Partially coupled in the Atlantic. Fully coupled elsewhere</td>
<td>Partially coupled in the Indo-Pacific. Fully coupled elsewhere</td>
</tr>
</tbody>
</table>

Figure 1 Zonal-meridional mask showing the partially coupled regions of the ocean component with prescribed momentum flux for TAU-PAC (25°S-60°N, left) and TAU-ATL (25°S-65°N, right) in black. White regions are where the model is fully coupled. The tapering regions (5 degrees latitude) are given by the gray shading.
Figure 2 Correlation between detrended HadISST data and SST from the ensemble means of (a) CNTRL, (b) TAU, (c) TAU-PAC and (d) TAU-ATL, for the period 1901-2000. Black dots indicate significant correlations at the 5% level for the effective degrees of freedom.

Figure 3 11-year filtered Northern Hemisphere surface temperature anomalies for GISTEMP (black line) and the ensemble simulations. The thick lines mark when the ensembles are significantly different from CNTRL at the 5% level, from a t-test based on the ensemble member spread.
Figure 4 11-year filtered zonal average surface temperature anomalies for GISTEMP and the 4 experiments.

Figure 5 11-year filtered annual Arctic (70-90°N) surface temperature for GISTEMP (black line) and the ensemble simulations. The thick lines mark when the ensembles are significantly different from CNTRL at the 5% level, from a t-test based on the ensemble member spread.
Figure 6 Significant (5% level) annual trends for the period 1910-1940 over the Arctic for GISTEMP (upper panel) and surface temperature (K/decade) in color and 500 hPa geopotential height (m/decade) in contours of the ensemble means.
Figure 7 11-year filtered Arctic (70-90°N) surface temperature for (a) the Atlantic side (90°W-90°E) and (b) the Pacific side (90°E-90°W) for GISTEMP (black line) and the ensemble simulations with same colors as in Figure 3 and 5. The thick lines mark when the ensembles are significantly different from CNTRL at the 5% level, from a t-test based on the ensemble member spread.

Figure 8 Annual surface temperature trend in K/decade for GISTEMP (a) and the ensemble means (b-e) for the period 1910-1940. Colors indicate regions with significant trends at the 5% level.
Figure 9  Annual SLP trend in hPa/decade for (a) HadSLP2 data and (b-d) the ensemble means for the period 1910-1940 north of 30°N. Filled contours indicate regions with significant trends at the 5% level.
Figure 10 Annual mean trend of geopotential height at 200hP (m/decade) from (a) 20CR and (b-e) the ensemble means for the period 1910-1940. Colors indicate regions with significant trends at the 5% level.
Figure 11 PDO-index for HadISST (black line) and the ensemble simulations. Thin lines are ensemble members. Thick lines are the ensemble means.

Figure 12 AMV-index for HadISST (black line) and the ensemble simulations. Thin lines are individual ensemble members. Thick lines are the ensemble means.
Supplementary figures

**Figure S1** Boreal winter mean (December-January-February) Nino3-index (5°N-5°S, 150°W-90°W) for HadISST data (black), and the ensemble means. The correlation with the observed Nino3-index is 0.74 (p<0.0001) for both TAU and TAU-PAC, 0.10 for TAU-ATL and -0.20 for CNTRL.

**Figure S2** 11-year filtered annual surface temperature anomalies for GISTEMP (black line) and the ensemble simulations for the Southern Hemisphere (90°S-0).

**Figure S3** Regression of the first principle component of the low-frequency filtered prescribed momentum flux curl anomalies onto low-frequency filtered global SST in TAU. Black dots indicate significant regions at a 5% level.