On Anomalous Ocean Heat Transport toward the Arctic and Associated Climate Predictability

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ABSTRACT

A potential for climate predictability is rooted in anomalous ocean heat transport and its consequent influence on the atmosphere above. Here we assess the propagation, drivers, and atmospheric impact of heat anomalies within the northernmost limb of the Atlantic meridional overturning circulation using a multi-century climate model simulation. Consistent with observation-based inferences, simulated heat anomalies propagate from the eastern subpolar North Atlantic, into, and through the Nordic Seas. The dominant time scale of associated climate variability in the northern seas is 14 years, including that of observed sea surface temperature and modeled ocean heat content, air–sea heat flux, and surface air temperature. A heat budget analysis reveals that simulated ocean heat content anomalies are driven by poleward ocean heat transport, primarily related to variable volume transport. The ocean’s influence on the atmosphere, and hence regional climate, is manifested in the model by anomalous ocean heat convergence driving subsequent changes in surface heat fluxes and surface air temperature. The documented northward propagation of thermohaline anomalies in the northern seas and their consequent imprint on the regional atmosphere – including the existence of a common decadal time scale of variability – detail a key aspect of eventual climate predictability.
1. Introduction

The Atlantic region exhibits distinct interannual to multidecadal variability (Deser and Blackmon 1993; Kushnir 1994; Frankcombe et al. 2010; Williams et al. 2014), reflected in upper-ocean thermohaline anomalies that propagate persistently through the North Atlantic Ocean and Nordic Seas toward the Arctic (Sutton and Allen 1997; Polyakov et al. 2005; Holliday et al. 2008). Decadal variability in ocean temperature plays an important role in the marine climate system (e.g., Drinkwater et al. 2014), influencing marine life from primary production to cod stocks (Helland-Hansen and Nansen 1909; Hátn et al. 2009). Ocean heat anomalies also play an important role in Arctic sea ice variability (Francis and Hunter 2007; Árthun et al. 2012; Onarheim et al. 2014; Carmack et al. 2015), which in turn could influence weather conditions and climate (e.g., Screen et al. 2013; Vihma 2014).

Anomalous ocean heat can extend its influence beyond the marine climate by being imprinted on the atmosphere (Rhines et al. 2008; Farneti and Vallis 2011; Gulev et al. 2013; Schlictholz 2013), acting to increase the persistence of atmospheric circulation anomalies and, hence, provide predictability of atmospheric variability and continental climate (e.g., Sutton and Hodson 2005). This, however, requires that oceanic variability is communicated to the atmosphere through surface heat fluxes. Understanding the mechanisms and time scales involved in the propagation of ocean heat anomalies and how they interact with the atmosphere is thus a prerequisite for skillful climate predictions for the North Atlantic/Arctic sector (Latif and Keenlyside 2011; Meehl et al. 2014).

The flow of warm, saline Atlantic waters toward higher latitudes takes place with the North Atlantic Current and its poleward extension, the Norwegian Atlantic Current (NwAC; Fig. 1a). In the Nordic Seas, the NwAC consists of two branches; a western branch enters the Nordic Seas over the Faroe–Iceland Ridge and is topographically guided northward along the front between
the Arctic and Atlantic waters, while a warmer and more saline eastern branch inflows through
the Faroe–Shetland Channel and continues north as a near-barotropic shelf-edge current (Orvik
et al. 2001). Upon reaching the western boundary of the Barents Sea the eastern branch of the
NwAC bifurcates flowing eastward into the Barents Sea, while the northward flow converges with
the western NwAC and continues toward the Fram Strait as the West Spitsbergen Current. Part of
the West Spitsbergen Current continues north into the Arctic Ocean (Rudels et al. 1999), but the
majority of the current recirculates westward in the Fram Strait (Bourke et al. 1988) and joins the
southward flowing deeper branch of the East Greenland Current en route to the Denmark Strait,
thus forming a cyclonic loop within the Nordic Seas. While traversing the periphery of the Nordic
Seas and the Arctic Ocean, the warm and saline Atlantic water is gradually transformed into a
colder and fresher outflow as a result of oceanic heat loss and freshwater input (Mauritzen 1996;
Rudels et al. 1999). Following Eldevik et al. (2014), the three regions connected by the NwAC
(Fig. 1a) – the northern North Atlantic, the Nordic Seas, and the Arctic Ocean – will hereafter be
jointly referred to as the northern seas.

Temperature anomalies have been observed to propagate northwards from the eastern subpolar
North Atlantic along the path of the North Atlantic Current and NwAC (Furevik 2000; Holliday
et al. 2008; Chepurin and Carton 2012; Yashayaev and Seidov 2015). In the northern seas, anom-
alias travel from the Greenland–Scotland Ridge to the west coast of Svalbard in approximately 1–3
years (Dickson et al. 1988; Eldevik et al. 2009). This corresponds to a propagation speed of 2–5
cm s$^{-1}$, which is an order of magnitude less than the typical current speed of the NwAC (Orvik
et al. 2001). Both anomalous air–sea heat fluxes due to changes in the large-scale atmospheric cir-

culation (Furevik and Nilsen 2005) and changing composition and strength of ocean currents have
been suggested to generate temperature anomalies in the northern seas (e.g., Furevik 2001; Carton
et al. 2011; Mork et al. 2014). Specifically, Mork et al. (2014) found that heat fluxes explain about
half of the observed interannual heat content variability, but that the fraction varies considerably
in time. Carton et al. (2011), on the other hand, found that surface heat flux variations in some
cases act to reinforce anomalies, but that the contribution was too small to explain the concomitant
changes in ocean heat storage. It is in most cases, however, not possible to construct a closed
observation-based heat budget because of sparse data coverage, especially in terms of ocean cur-
rent measurements. The relative importance of ocean and atmosphere in modifying ocean heat
anomalies can therefore not be fully distinguished from observations. Heat flux reanalysis prod-
ucts also partly disagree, making it sometimes problematic to compare with changes in observed
hydrography (Carton et al. 2011).

The purpose of this paper is twofold: To disentangle the contributions from ocean circulation
and air–sea exchange in the propagation of ocean heat anomalies from the North Atlantic toward
the Arctic, and to assess the potentially predictable relation between anomalous ocean heat and
climate in the northern seas region. To this end, a 500-year control simulation with the Bergen
Climate Model is used (Otterå et al. 2009). The model analysis is aided by historical sea surface
temperatures (HadISST; Rayner et al. 2003).

The model and the observations are introduced in section 2. In section 3 we evaluate the model
performance for the northern seas. The propagation and drivers of anomalies are then analyzed
in section 4 and section 5, while the link to upstream variability in the subpolar North Atlantic
is discussed in section 6. The atmospheric imprint of ocean heat anomalies and the identified
characteristic time scale of oceanic and atmospheric variability are discussed in section 7. Finally,
the main conclusions and implications are presented in section 8.
2. Data and Methods

a. Bergen Climate Model

This study uses a 500-year pre-industrial control simulation from the Bergen Climate Model (BCM), a fully coupled atmosphere-ocean-ice general circulation model. A general description of the model is given by Furevik et al. (2003), while the model run used is described in Otterå et al. (2009). Only a short summary is given here.

The ocean component of BCM is a modified version of the Miami Isopycnic Coordinate Ocean Model (MICOM; Bleck et al. 1992). The version used for this model run uses potential density with reference pressure at 2000 dbar as vertical coordinates (\(\sigma_2\) coordinates). The ocean model consists of 34 isopycnic layers, ranging from \(\sigma_2 = 30.119 \text{ kg m}^{-3}\) to \(\sigma_2 = 37.800 \text{ kg m}^{-3}\), below a non-isopycnic mixed layer. The mixed layer depth is calculated from the turbulent kinetic energy balance of a Kraus-Turner type one-dimensional mixed layer model (Gaspar 1988), with modifications detailed in Medhaug et al. (2012). The horizontal grid resolution is 2.4° longitude \(\times\) 0.8° latitude at the equator, becoming more isotropic with increasing latitude. In the northern seas the horizontal resolution ranges from 70–100 km. MICOM is coupled to a multi-category dynamic-thermodynamic sea ice model, GELATO (Salas-Mélia 2002). The atmospheric component is ARPEGE-CLIMAT3 (Déqué et al. 1994), a low-top spectral model with a horizontal resolution of \(~2.8^\circ\) and 31 vertical levels. Fluxes of mass, energy, and/or momentum are calculated in ARPEGE and communicated to the ocean via the OASIS (Terray and Thual 1995) coupler. External forcing from, e.g., solar insolation and greenhouse gases, is set to constant pre-industrial values. No flux corrections are applied, allowing the model to freely develop its own climatology. The initial conditions for the pre-industrial control simulation are obtained from the end of a 500-
year spin-up integration (detailed in Otterâ et al. 2009), after which the model is run for 600 years. Here we use the last 500 years of the simulation.

b. Methods

Ocean heat anomalies are assessed as follows. The heat content is calculated as:

\[ H(x, y, t) = \rho_w c_p \int_{\eta}^{D} [T(x, y, t) - T_{\text{ref}}] \, dz, \]  

where \( T \) is temperature, \( \rho_w \) is density of seawater, and \( c_p \) is the specific heat capacity of seawater.

The heat content is calculated between the free surface \( \eta \) and the full depth of the ocean, \( D \), using a reference temperature \( T_{\text{ref}} = -2^\circ C \), although we note that heat content anomalies (relative to the local mean) presented herein are practically insensitive to the specific reference temperature (not shown). We also note that an ocean heat budget calculated from the heat convergence of a closed mass budget (section 5) is independent of \( T_{\text{ref}} \). Monthly anomalies at each grid point are then obtained by subtracting the respective mean seasonal cycle. Unless stated otherwise time series are then low-pass filtered with a third-order Butterworth filter with a cut-off period of 3 years to emphasize interannual to decadal variability.

Statistical significance is assessed using a two-tailed Student \( t \)-test, adjusted for serial autocorrelation (Chelton 1983). All correlations given in the text are significant at the 95% confidence level.

c. Complex principal component analysis

Propagating phenomena can be identified and objectively analyzed from a complex principal component (CPC) analysis which detects traveling waves in the input time series. A full description of the procedure is given by Horel (1984), and only a short summary is presented here. First, a complex dataset \( f(x, t) \) is formed from the original data (by rotating its Fourier components \( \pi/2 \)).
Complex eigenvectors are then computed from the cross-covariance matrix derived from the complex dataset. From the covariance matrix, the CPCs $[P_n(t)]$ and the complex empirical orthogonal functions [CEOFs; $F_n(x)$] are calculated. The (complex) dataset can thus be represented as a sum of the contribution from the $N$ principal components:

$$f(x,t) = \sum_{n=1}^{N} P_n(t) F_n^*(x),$$

where $x$ denotes spatial position and $t$ is time. The asterisk denotes complex conjugation. The elements of the CPC time series can furthermore be written in the form of an amplitude $a_n$ and a phase $\phi_n$: $P_n(t) = a_n(t) e^{i\phi_n(t)}$. The importance of each CEOF (mode) is defined as the proportion of variance explained by each principal component.

**d. Observed SST**

To corroborate the model analysis we use sea surface temperature (SST) data from the Hadley Centre (HadISST; Rayner et al. 2003) covering the period between 1870 and 2013. These data have a spatial resolution of 1° longitude by 1° latitude and monthly temporal resolution. We will only consider winter (December–April) SST as it represents the upper-ocean heat content as a result of a deep winter mixed layer (Nilsen and Falck 2006). The observation-based analysis is furthermore restricted to the southern Norwegian Sea to avoid the potential influence of sea ice. A good agreement in terms of interannual to decadal variability has been found between HadISST and data from standard hydrographic sections in the northern seas (Hughes et al. 2009).

**3. Model performance in the northern seas**

To assess the propagation of ocean heat anomalies in the northern seas it is essential that the model of choice is able to adequately represent the northward flow of Atlantic water and the gradual transformation into dense overflow water as it circulates the periphery of the Nordic Seas.
Previous applications of the Bergen Climate Model in the northern seas (e.g., Otterå et al. 2010; Langehaug et al. 2012a,b; Medhaug et al. 2012; Lohmann et al. 2014) have found that the model realistically simulates the structure and the mean poleward heat transport of the NwAC, as well as the dense overflow and fresh surface waters in the Denmark Strait. Notably, the thermohaline contrast between these three water masses occupying the Greenland–Scotland ridge is consistent with observations, although the model hydrography is skewed toward warmer and more saline properties. For the Atlantic inflow specifically, the model is $\sim 2^\circ$C warmer and $\sim 0.5$ saltier than observations (cf. Fig. 7 in Langehaug et al. 2012a). The associated modeled volume transport into the Nordic Seas is 7.4 Sv for the NwAC, 2.1 Sv in the East Greenland Current, and 5.7 Sv of overflow water, which is in good agreement with observational estimates of 8.5 Sv, 0.4–2.1 Sv, and 6.4 Sv respectively (see Langehaug et al. 2012a and references therein). The model also captures the bifurcation at the western boundary of the Barents Sea (Fig. 1b), with a heat transport into the Barents Sea (61 TW; $1 \text{ TW} = 10^{12} \text{ J s}^{-1}$; Medhaug et al. 2012) which is close to that observed (Årthun et al. 2012). The realistic model transports and properties of inflowing and outflowing waters point to an accurate modification of water masses within the Nordic Seas. This is corroborated by Langehaug et al. (2012b) who found a realistic structure of surface-forced water mass transformation (diagnosed from surface buoyancy fluxes) along the path of the North Atlantic Current in the model.

The observed and simulated winter sea ice extent is shown in Figure 1. The model ice cover is in good agreement with observations in the Norwegian Sea and western Barents Sea, while it is generally larger than the observed in the Greenland Sea. However, as noted by Smedsrud et al. (2013), a more extensive ice cover in the model is reasonable as the simulation uses constant pre-industrial external forcing, and therefore does not include recent sea ice decline in the Arctic. The simulated minimum and maximum winter sea ice edge is furthermore in agreement with the
observed minimum and maximum (Smedsrud et al. 2013), i.e., the model variability is within the observed range.

Admittedly, the Bergen Climate Model and other global climate models are unable to resolve the smaller scale features of ocean circulation, e.g., mesoscale ocean eddies and narrow boundary currents. However, multidecadal variability of the coupled atmosphere–ocean system can only be studied using relatively coarse climate models, as multicentury simulations with eddy-resolving grid resolution are generally not available. Fully coupled climate models are nevertheless valuable and necessary tools for assessing climate variability on decadal time scales and beyond.

4. Propagation of simulated ocean heat anomalies

To assess the properties and modification of modeled ocean heat anomalies in the northern seas we first need to determine the path of propagation. Based on the mean circulation (illustrated by the barotropic stream lines in Figure 2) and extent of the simulated NwAC (Fig. 1b), 11 stations (St1–11) have been defined that capture the mean propagation. This includes one station in the North Atlantic, corresponding to the Rockall Trough, one at the Greenland–Scotland ridge, and nine stations downstream within the Nordic Seas. The grid points included in each station are shown as circles in Figure 2. The lagged correlation between adjacent stations is generally high (>0.7; lags vary, but are typically around 6 months) for both heat (Fig. 2) and salt content (not shown), except for lower values between St2 and St3, and St9–11. The former could be a result of both variable communication between the North Atlantic and Nordic Seas (e.g., Hátún et al. 2005) and hydrographic variability internal to the Nordic Seas (e.g., Mork et al. 2014), while the less coherent signal between St9 and St11 could be a result of the branch of Atlantic water entering the Nordic Seas west of Iceland (Fig. 1b; Langehaug et al. 2012a) or branching of the southward flow in the south Greenland Sea (Mauritzen 1996).
The time evolution of ocean heat anomalies along the defined path is shown in Figure 3a. Propagating warm and cold anomalies are evident throughout the time period. Heat anomalies with a depth-averaged standard deviation of $1.4 \times 10^{19}$ J (corresponding to $0.5^\circ$C) are associated with salinity anomalies of 0.05, warm conditions being accompanied by higher salinities (Fig. 3a), i.e., anomalies being largely density compensated. Elevated variability is found in the southern end of the Norwegian Sea (St2–3) and in the area between Norway and Svalbard (St6–7). This spatial pattern is consistent with that inferred from observed Norwegian Sea heat content variability (Skagseth and Mork 2012). The observed anomalies discussed by Furevik (2001) also showed the largest amplitude in the Sørkapp section at $76.5^\circ$N (approximate position of St7). The along-path evolution (relative to the local mean) is determined by the concomitant anomalous forcing (ocean and atmosphere) within the northern seas. A northward strengthening of an anomaly can for instance be explained by anomalously low surface heat loss or by an increased advection speed (Furevik 2001).

In the North Atlantic and southern Norwegian Sea (St1–5) the magnitude of ocean heat content variability is largest between 300 m and 500 m depth (Fig. 4). The stations further downstream (St6–9) show maximum variability deeper in the water column (although less pronounced at St8–9), reflecting the along path modification and deepening of the Atlantic water. The marked change between St7 and St8 relates to the boundary between the ice-free Norwegian Sea and the seasonally ice covered Fram Strait and Greenland Sea (Fig. 1b), i.e., the northward extent of the Atlantic domain. The associated temperature variability is approximately $0.6–1.0^\circ$C in the Norwegian Sea, while it is smaller ($<0.4^\circ$C) in the Greenland Sea and in the Rockall Trough. The surface intensified variability at St6 and St7 is most likely related to large surface heat loss [both modeled and observation-based; Langehaug et al. (2012b)], whereas the large surface variations at St11 is caused by the Atlantic inflow west of Iceland. The depth of maximum heat content variability at...
St1–6 corresponds the average depth of the winter mixed layer (Fig. 4). Noting that the depth of the winter mixed layer in the Norwegian Sea reflects the base of the Atlantic layer (Nilsen and Falck 2006), this suggests that heat content changes are largely driven by changes in the layer thickness and, hence, volume of Atlantic water (Sandø et al. 2012). This will be elucidated further in the next section.

Propagating phenomena can be objectively assessed from a complex principal component analysis, which detects traveling waves in the input time series and orders the dataset into modes of phase propagation in space and time according to the variance explained (Section 2). The leading mode of phase propagation of along-path heat anomalies (Fig. 3b) explains 50% of the total variance in the full dataset (Fig. 3a) and is well separated from the second mode (accounting for 21% of the variance). The phase angle of the leading mode increases with increasing station number (Fig. 3c), which implies that the simulated heat anomalies predominantly travel along the rim of the basin in the direction of the mean current, consistent with observation-based inferences (Holli-day et al. 2008; Eldevik et al. 2009). The 500-year time period consists of 31 complete cycles with a phase propagation that is rather constant in time (Fig. 3d). This yields a period of 16 years.

The circulation from St1 to St11 constitutes just over half a cycle which, with a travel distance of about 4800 km, implies that the speed of modeled anomalies is on average 2 cm s\(^{-1}\), an estimate which is in reasonable agreement with observations (Furevik 2000; Polyakov et al. 2005; Holliday et al. 2008; Chepurin and Carton 2012).

The representativeness of the time scale associated with propagating anomalies obtained from the complex principal component analysis compared with the full variance can be evaluated by a frequency analysis of the anomalous heat content at individual stations (note that no filter is applied in the frequency analysis). Heat anomalies in the northern North Atlantic (St1) and the Norwegian Sea (St3) both have a significant (95% confidence level) characteristic time scale of
14 years (Fig. 5a). The 14-year time scale is also clearly identifiable for salinity (Fig. 5b). All Atlantic-dominated stations (St1–7; cf. Fig. 4) as well as downstream St8 and St9 show significant power on this time scale. As mentioned above, the weaker signature of a propagating signal at St10 and St11 is most likely a result of thermohaline anomalies exiting the Nordic Seas both west (St11) and east of Iceland (Mauritzen 1996; Eldevik et al. 2009). Significant interdecadal variability is also found in the observation-based HadISST winter sea surface temperatures between 1870 and 2013 for the same region (0.5–17.5°E, 60.5–71.5°N; Fig. 5c), increasing the confidence in the model’s ability to simulate climate variability in the northern seas. The modeled ocean heat content also displays significant multidecadal variations (∼40–50 years). Variability on this time scale will, however, not be addressed here and the reader is referred to e.g., Frankcombe et al. (2010) and references therein for a discussion on mechanisms for multidecadal variability in the North Atlantic.

5. Heat budget for the Norwegian Atlantic Current

To assess the relative roles of ocean advection and air–sea fluxes in driving ocean heat anomalies, and how anomalous ocean heat might imprint on the atmosphere, the depth-integrated heat budget for the Norwegian Sea (Fig. 6) is now assessed in particular. The chosen area corresponds to the Atlantic dominated eastern Nordic Seas (Fig. 1) where the heat content variability is highest (Fig. 6) and interaction with the atmosphere is strongest (Langehaug et al. 2012b). The heat budget area is also similar to that used in the observation-based heat content analysis by Carton et al. (2011) and Mork et al. (2014). Although heat content variability in the whole water column is considered, changes in Norwegian Sea heat content predominantly reflect variability within the well-mixed Atlantic layer ($r = 0.72$; Nilsen and Falck 2006; Chepurin and Carton 2012) which is in contact with the atmosphere.
Changes in heat content within a control volume occur as a result of the imbalance between the surface heat flux (area \(A\)) and advective and diffusive heat transports through the vertical boundaries (area \(S\)):

\[
\frac{\partial H}{\partial t} = \rho_w c_p \int_S vT dS - \int_A q dA + Q_{\text{res}},
\]

where \(H_t\) is the heat content tendency, \(v\) is the cross-sectional velocity into the control volume used to calculate the heat transport convergence (\(Q_{\text{adv}}\)), \(q\) is the net ocean–atmosphere heat flux (the sum of both turbulent and radiative components) which integrated over an area \(A\) yields \(Q_s\), and \(Q_{\text{res}}\) is a residual term representing the ocean heat transport into the domain resulting from parameterized diapycnal mixing and lateral turbulent mixing (Otterå et al. 2009). Note that positive surface heat fluxes indicate ocean heat loss. In Figure 6a the time-varying heat budget components are plotted as anomalies (for presentation purposes only the first 100 years are plotted, while all calculations are based on the full 500-year time series), calculated as previously described in relation to Eq. (1).

The heat content rate of change is strongly associated with ocean heat transport convergence, both in terms of variability \((r = 0.70)\) and amplitude. The larger contribution from oceanic variability in driving heat content change is generally the case within the northern seas (Fig. 7) and especially along the path of the NwAC, whereas air–sea fluxes are relatively more important in parts of the Greenland Sea and in the northwestern Barents Sea.

The advective heat budget for the Norwegian Sea is in turn dominated by anomalous heat transport between Iceland and Scotland \((r = 0.82)\), i.e., the Atlantic inflow \((HT_{\text{aw}}; \text{Fig. 6b})\). The heat transport anomalies associated with the Atlantic inflow lead the northern \((HT_{\text{wsc}})\) and eastern \((HT_{\text{bs}})\) outflow by about 2.5 years, consistent with the calculated speed of anomalies. Heat transport anomalies \((HT' = \rho_w c_p \int_S (vT') dS)\) can occur as a result of changes in advection speed \((v'T)\) or temperature \((\bar{T}')\), or through eddy fluxes \((v'T')\). The respective contributions of these terms to
the Iceland–Scotland heat transport are shown in Figure 8. The variability of $HT_{aw}$ is dominated by $v'\bar{T}$, i.e., velocity and thus volume transport fluctuations rather than temperature fluctuations drive heat transport anomalies into the area. The dominant role of ocean advection on heat storage rates on interannual to decadal time scales agrees with recent findings from the subpolar North Atlantic (Buckley et al. 2014; Williams et al. 2014).

Heat anomalies in the Norwegian Sea (Fig. 6), rooted in the ocean, are furthermore found to force changes in air–sea fluxes. The correlation between ocean heat transport convergence and surface heat loss is 0.66, i.e., anomalously high ocean heat transport corresponds to enhanced heat loss to the atmosphere. Consistent with variable air–sea exchange associated with ocean heat anomalies, the surface heat flux and surface air temperature (SAT) within the Norwegian Sea (evaluated at grid points shown in Fig. 6) also have the same decadal-scale oscillation of 14 years (Fig. 9a). The atmospheric temperature anomalies covary with ocean heat content ($r = 0.73$), leading to a reduced thermal contrast between ocean and atmosphere and thus contributing to weaker air–sea fluxes and reduced damping of ocean heat anomalies [$r(\text{SAT}, Q_s) = -0.47$]. Changes in surface air temperature lag variations in the Atlantic inflow ($HT_{aw}$) by 2–3 years ($r = 0.57$). The potential predictability of atmospheric variability from ocean heat anomalies is further discussed in section 7.

6. Source of northern seas heat anomalies

The temporal development of heat content anomalies in the northern seas (Fig. 3) implies a propagating signal. The poleward progression of thermohaline anomalies – heat and freshwater – is also a robust finding in observations and in other ocean and climate models (e.g., Dickson et al. 1988; Hansen and Bezdek 1996; Krahmann et al. 2001; Holliday et al. 2008; Chepurin and Carton 2012; Glessmer et al. 2014). There is nevertheless no complete mechanistic understanding of
the driving mechanisms of these anomalies, including the roles of ocean dynamics and stochastic atmospheric forcing (see review by Liu 2012). In support of the latter, a number of studies have related low-frequency temperature variability in the North Atlantic to atmospheric variability associated with the winter North Atlantic Oscillation (NAO; Hurrell 1995). Visbeck et al. (1998) and Krahmann et al. (2001) demonstrated how the formation and propagation of temperature anomalies along the pathway of the North Atlantic Current can be obtained by temporal changes in NAO-like wind forcing. The propagation was found to be a result of both advection of existing temperature anomalies by the mean ocean currents and locally generated anomalies from spatial variations in the external forcing. Simple advection of coherent temperature anomalies through the North Atlantic is also not supported by recent drifter studies (e.g., Burkholder and Lozier 2014), showing no direct advective pathway of anomalous heat between the subtropical and subpolar region.

Variable (NAO-like) atmospheric forcing can also induce upper-ocean temperature anomalies through modulation of the North Atlantic Ocean circulation and subpolar gyre (SPG) strength, driving changes in poleward heat transport (Czaja and Marshall 2001; Eden and Jung 2001; Lohmann et al. 2009). This has also been shown for the Bergen Climate Model (Langehaug et al. 2012a; Medhaug et al. 2012). It has previously been demonstrated both from observations (e.g., Hátún et al. 2005; Yashayaev and Seidov 2015) and modeling studies (e.g., Jungclaus et al. 2014) that variability in the amount and temperature of Atlantic water flowing northward across the Greenland–Scotland ridge are driven, in part, by the strength of the SPG. In the Bergen Climate Model this is resonated in enhanced spectral power at the same frequencies for the modeled heat transport by the NwAC \(H_{\text{aw}}\) and the SPG strength (calculated from the barotropic stream-function in the subpolar region), including the 14-year periodicity found in ocean heat anomalies within the northern seas (Fig. 9b). A strengthening of the subpolar gyre precede heat and salt content changes in the Norwegian Sea by 1 year, \(r = 0.46\) and \(r = 0.60\), respectively. Our results
thus support a close coupling between the subpolar North Atlantic and climate variability in the northern seas through variations in poleward heat transport.

### 7. Oceanic forcing of atmospheric variability

The identified persistent northward advection of anomalous ocean heat and the consequent decadal changes in surface heat fluxes and surface air temperature (Fig. 6; Fig. 9a) suggest potential climate predictability in the northern seas region. To further demonstrate the predictive capability associated with a variable ocean heat transport, Figure 10 shows the two-year lagged response in northern seas surface heat fluxes (Fig. 10a) and surface air temperature (Fig. 10c) to changes in the Atlantic inflow based on linear regression. Inflow-driven heat fluxes of $5–20\; W\; m^{-2}$ (per standard deviation of heat transport) occur offshore of the Norwegian coast and further north in the marginal ice zone around Svalbard and in the Barents Sea, reflecting fluctuations in the zonal and meridional extent of the Atlantic domain, and hence sea ice extent (Fig. 10b), respectively. The air temperature response is, on the other hand, pronounced over large parts of the northern seas.

The magnitude of the atmospheric response to ocean heat anomalies varies from $>0.5^\circ C$ in the marginal ice zone to $0.1–0.3^\circ C$ in the southern Norwegian Sea (Fig. 10c), values being similar to temperature anomalies in the ocean (Fig. 4). There is also a significant atmospheric response over land ($0.1–0.4^\circ C$), in agreement with Norwegian climate (air temperature) reflecting decadal temperature variability in the Norwegian Sea (e.g., Eldevik et al. 2014), and, more broadly, European continental climate reflecting North Atlantic SSTs (e.g., Sutton and Hodson 2005).

Understanding the interaction between the ocean and the atmosphere is a prerequisite for understanding and predicting climate variability. To what extent anomalous ocean heat leads to atmospheric circulation changes is not addressed here. Modeling studies have generally considered that the amplitude of the atmospheric response to extratropical large-scale SST anomalies...
is modest compared with internal atmospheric variability (Kushnir et al. 2002), but this is still a matter of debate and might be both model (Omran et al. 2014; Smirnov et al. 2015) and time scale dependent (Gulev et al. 2013; Sheldon and Czaja 2014). Ocean heat anomalies in the northern seas can in any case yield a significant regional atmospheric response. In line with our results van der Swaluw et al. (2007), also using data from a pre-industrial control run, showed that anomalous heat transport by the NwAC forces the atmosphere by increased heat fluxes in the marginal ice zone (>70°N; cf. our Fig. 10b). The atmospheric response to increased heat transport was associated with a cyclonic pressure anomaly and decreased atmospheric heat transport by baroclinic eddies as a result of a decreased poleward temperature gradient in the atmosphere. Similar mechanisms and impact were also identified by Schlichtholz (2013) based on observational data from the northern seas. The regionally confined anomalous atmospheric circulation in response to decadal-scale ocean-driven sea ice variability in the northern seas, and in particular in the Barents Sea, can also drive larger-scale surface climate variability (e.g., Semenov et al. 2010; Liptak and Strong 2014).

In support of decadal ice–ocean interaction in the Barents Sea, the modeled winter (December–April) sea ice cover in the Barents Sea (15–60°E, 70–81°N) has a similar spectrum to that of climate variability in the northern seas (Fig. 5; Fig. 9b), including a dominant time scale of 14 years (not shown). The correlation between the low-pass filtered heat transport into the Barents Sea ($HT_{bs}$) and sea ice extent in the Barents Sea is -0.77, with a time lag of 1–2 years in agreement with observations (Årthun et al. 2012; Onarheim et al. 2015). The decadal-scale oscillation furthermore agrees with Venegas and Mysak (2000) and Vinje (2001) who found fluctuations in the observed Barents Sea ice extent with a time scale of 16–20 years and 12–14 years, respectively, related to a variable NwAC.

Identifying a time scale associated with northward propagating ocean heat anomalies from the subpolar North Atlantic toward the Arctic (Fig. 5; Fig. 9b) and their consequent interaction with
the atmosphere (Fig. 9a; Fig. 10) is essential in terms of potential climate prediction. A recent model study by Escudier et al. (2013) found that a 20-year coupled mode of atmosphere–ice–ocean variability may exist in the subpolar North Atlantic in which the propagation of thermohaline anomalies from the subpolar gyre interacts with the atmosphere in the northern seas to eventually produce anomalies of the opposite sign in the Labrador Sea. Results demonstrated herein further supports a coupled mode of variability in the northern seas associated with the propagation and atmospheric imprint of ocean heat anomalies. The different time scale in the two models is likely related to the stronger northward heat transport in the Bergen Climate Model (Langehaug et al. 2012b) as the time scale of the cycle is set by the propagation of anomalies along the rim of the Nordic Seas. The robustness of the time scale identified herein needs to be further assessed, but we reiterate that the characteristic 14-year time scale of climate variability in the northern seas is supported by observed SST fluctuations in the Norwegian Sea (Fig. 5c).

8. Conclusions

Interannual to decadal-scale ocean heat anomalies associated with the northern limb of the Atlantic meridional overturning circulation propagate persistently toward the Arctic (e.g., Holliday et al. 2008). This poleward propagation of anomalous ocean heat is commonly understood to be a primary source for climate predictability (e.g., Latif and Keenlyside 2011). Here, we have used a 500-year control simulation from the fully coupled Bergen Climate Model (BCM), aided by observed sea surface temperatures (HadISST), to assess the propagation and drivers of ocean heat anomalies in the northern seas (northern North Atlantic, Nordic Seas, and Arctic Ocean), and to what extent these anomalies imprint on the atmosphere.

Ocean heat anomalies are found to propagate from the eastern subpolar North Atlantic, into, and along the rim of the Nordic Seas with a speed of 2 cm s\(^{-1}\) (Fig. 3). The characteristic time
scale of variability is 14 years, which is also that of observed sea surface temperature variability in the Norwegian Sea during the last century (Fig. 5). The relative roles of ocean and atmosphere in driving ocean heat anomalies are assessed by constructing a depth integrated heat budget for an area covering the Atlantic domain of the Nordic Seas, i.e., the Norwegian Sea. Changes in heat content are found to be caused mainly by anomalous ocean heat transport convergence (Fig. 6). Variations in ocean heat convergence largely originate in the inflow from the Atlantic proper, and a temporal decomposition of the Atlantic heat transport shows that volume transport anomalies dominate (Fig. 8). Simulated ocean heat anomalies in the northern seas are thus driven mainly by changes in the strength of the northward flowing Atlantic water. A similar decadal-scale oscillation in the strength of the subpolar gyre (Fig. 9b) further supports the close coupling, observed and modeled, between the subpolar North Atlantic and Nordic Seas–Arctic Ocean (e.g., Hátún et al. 2005; Glessmer et al. 2014; Jungclaus et al. 2014).

A potentially predictable relation between anomalous ocean heat and climate in the northern seas region is furthermore identified. Ocean heat anomalies in the northern seas are reflected in regional sea ice extent and found to influence the atmosphere by driving changes in surface air temperatures through anomalous air–sea fluxes (Fig. 10). The interaction with the atmosphere is also most pronounced on a 14-year time scale. The identified time scale of climate variability, manifested both in anomalous ocean heat transport and its consequent atmospheric response, provides encouraging evidence for climate predictability rooted in the northern seas.

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model data and Helene R. Langehaug for providing the SPG index. We also thank Tore Furevik and three anonymous reviewers for valuable comments which improved the manuscript.

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Fig. 1. a) Observed and b) modeled winter (December–April) upper-ocean temperature (shading) and sea ice extent (white line; defined where the sea ice concentration is 15%) in the northern seas. Observations are from HadISST (Rayner et al. 2003). The arrows indicate the main features of the near-surface circulation (NwAC: Norwegian Atlantic Current; WSC: West Spitsbergen Current; EGC: East Greenland Current). Modeled temperature and velocities are averages within the surface mixed layer (note the different velocity scales). Isobaths are given every 1000 m (thin gray lines) and for 500 m depth (black line) which roughly marks the continental slopes. DS: Denmark Strait; RT: Rockall Trough; Sh: Shetland.

Fig. 2. Barotropic streamlines showing the mean cyclonic circulation within the northern seas (thick black lines; plotted for -1.5 Sv, -2 Sv, and -3 Sv; 1 Sv ≡ 10$^6$ m$^3$ s$^{-1}$). Isobaths are given for 500 m, 2000 m, and 3000 m depth (gray lines). The colored circles show the grid points used to define the path of ocean heat anomalies, where the associated color indicates the maximum lagged correlation between adjacent stations, i.e., the correlation shown for St2 corresponds to r(St1,St2). Station numbers are assigned and used in the text.

Fig. 3. a) Temporal development of low-pass filtered heat content (shading; units in 10$^{19}$ J) and depth-averaged salinity anomalies (contours; plotted for 0 and ±0.05, dashed lines corresponding to a negative salinity anomaly) at defined stations (averages over grid points shown in Fig. 2). The thick dashed line corresponds to a propagation speed of 2 cm s$^{-1}$. b) The dominant mode of propagation derived from complex principal component analysis (units in 0.5 × std. deviation), explaining 50% of the total variance. c,d) Phase angle as a function of station number and time, respectively. Increasing phase angle with increasing distance (station) corresponds to a cyclonic propagation. The dashed line in c) corresponds to a propagation speed of 2 cm s$^{-1}$, whereas in d) it corresponds to constant phase propagation.

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