Causes of the Rapid Warming of the North Atlantic Ocean in the Mid-1990s

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ABSTRACT

In the mid-1990s, the subpolar gyre of the North Atlantic underwent a remarkable rapid warming, with sea surface temperatures increasing by around 1°C in just 2 yr. This rapid warming followed a prolonged positive phase of the North Atlantic Oscillation (NAO) but also coincided with an unusually negative NAO index in the winter of 1995/96. By comparing ocean analyses and carefully designed model experiments, it is shown that this rapid warming can be understood as a delayed response to the prolonged positive phase of the NAO and not simply an instantaneous response to the negative NAO index of 1995/96. Furthermore, it is inferred that the warming was partly caused by a surge and subsequent decline in the meridional overturning circulation and northward heat transport of the Atlantic Ocean. These results provide persuasive evidence of significant oceanic memory on multiannual time scales and are therefore encouraging for the prospects of developing skillful predictions.

1. Introduction

The global oceans have warmed since the mid-twentieth century (Levitus et al. 2009; Ishii and Kimoto 2009; Wijffels et al. 2008), largely consistent with the radiative forcing caused by greenhouse gases (Solomon et al. 2007). However, the world’s oceans have not warmed uniformly in space or time (Harrison and Carson 2007). Understanding the causes of the inhomogeneity is essential given the central role of ocean heat uptake for regulating the climate, the importance of understanding variability for the detection and attribution of climate change, and the impact of sea surface temperatures (SST) on the evolution of regional climates.

The Atlantic is one area of the global ocean that has exhibited significant multidecadal variability in sea surface temperatures (Kushnir 1994; Ting et al. 2009). In the last part of the twentieth century, the North Atlantic Ocean warmed significantly (Palmer and Haines 2009). In particular, the North Atlantic subpolar gyre (SPG) region has warmed substantially since the mid-1990s, especially in the eastern SPG (Sarafanov et al. 2008; Marsh et al. 2008; Holliday et al. 2008; Grist et al. 2010; Reverdin 2010). The warming of the North Atlantic subpolar gyre SST and heat content is linked to a slowdown of the SPG circulation (Häkkinen and Rhines 2004) and has been implicated in decreased carbon uptake by the North Atlantic (Corbière et al. 2007; Schuster et al. 2009), significant changes in marine ecosystems (Hátún et al. 2009), changes in Atlantic Hurricane numbers (Smith et al. 2010), and increased melting of the Greenland ice sheet (Holland et al. 2008).

It is well established that the North Atlantic Oscillation (NAO; Hurrell 1995) is important for driving the variability in the North Atlantic. For example, Lozier et al. (2008) showed that the major changes in heat content in the North Atlantic Ocean between the periods 1950–70 and 1980–2000 could be understood as a response to decadal variations in the NAO. From the
1970s until the mid-1990s, a trend toward positive NAO was observed (Hurrell 1995). Generally speaking, positive NAO leads to stronger winds across the North Atlantic and hence acts to cool the surface of the subpolar North Atlantic Ocean through increased turbulent fluxes (Marshall et al. 2001; Visbeck et al. 2003). However, since the mid-1990s the NAO has been largely neutral. A warming of the North Atlantic could potentially be explained by the reduction of the NAO index after the early 1990s. But was this really the cause? Such a hypothesis assumes no significant role for nonlocal dynamical changes in the ocean.

Since the mid-1990s, significant changes in the surface currents of the North Atlantic have been observed (Flatau et al. 2003; Verbrugge and Reverdin 2003; Håkkinen and Rhines 2009). There is considerable evidence from model studies that dynamical changes should be expected on a variety of time scales in response to changes in the NAO. Changes in wind stress impact on Ekman transports, and changes in wind stress curl impact on the gyre transports (Eden and Willebrand 2001; Marshall et al. 2001). Ocean adjustments to wind stress curl changes involve the westward propagation of Rossby waves associated with anomalies in isopycnal depth and heat content (Schneider et al. 2002; Leadbetter et al. 2007).

There has also been much focus recently on changes in the spatial extent of the SPG. As the strength of the SPG declined in the mid-1990s, it also contracted (Bersch et al. 2007; Sarafanov et al. 2008), with a northwestward shift of the subpolar front, consistent with the modeling study of Hátún et al. (2005). However, there is uncertainty over the primary driver of the SPG circulation changes. Changes in wind stress are important for the strength of the gyre circulation (Eden and Willebrand 2001; Brauch and Gerdes 2005), but the strength of the SPG is also dependent on the buoyancy forcing (Bersch et al. 2007; Lohmann et al. 2009a). Håkkinen and Rhines (2004) and later Bersch et al. (2007) suggest that the reduction of the buoyancy-forced deep convection was important for the decline in the size and strength of the SPG since the mid-1990s. However, Herbaut and Houssais (2009) and Håkkinen et al. (2011) argue that wind forcing dominates changes in the eastern SPG. Despite the uncertainty, all these studies suggest that the warming of the SPG could be explained by a dynamical response (contraction and weakening) due to the decline in the NAO since the mid-1990s.

However, Lohmann et al. (2009b) suggest that the SPG would have weakened since 1995 without any decline in the NAO, albeit not as dramatically. They argue the decline in the SPG was driven by an increase in the northward heat transport of the ocean due to a lagged response to the extended positive phase of the NAO in the late 1980s and early 1990s. Models predict that a positive NAO leads to an increase in the Atlantic meridional overturning circulation (AMOC) through an increase in deep-water formation due to increased cooling in the North Atlantic (Delworth and Greatbatch 2000; Eden and Willebrand 2001; Dong and Sutton 2005; Lohmann et al. 2009a; Ortega et al. 2012). In the models, the acceleration of the AMOC typically leads to an increase in the northward heat transport and a warming of the North Atlantic Ocean, which lags the NAO by a number of years. Thus, the overturning and the associated heat transport could well have played a role in the warming.

To summarize, there are two general hypotheses for the warming of the North Atlantic in the mid-1990s. They are as follows:

1) The warming was caused by the decline in the NAO since 1995.
2) The warming was a delayed response to the high NAO index that preceded the warming.

One reason for understanding the relative importance of these mechanisms is that they have very different consequences for the predictability of the observed warming event. Given that the predictability of the NAO appears to be low beyond one or two months (Collins 2002; Müller et al. 2005), the first hypothesis implies low predictability of the warming, whereas the second hypothesis could imply high predictability.

This study will attempt to understand the relative importance of the hypothesized mechanisms that have been described by comparing model results with ocean observations. The paper is organized as follows: Section 2 describes the ocean analyses and model experiments. In section 3, a comparison of the observed and modeled changes in the Atlantic is presented, before a discussion of the important mechanisms involved is presented in section 4. A wider discussion of the results in the context of other studies is presented in section 5, and a summary and conclusions are in section 6.

2. Ocean analyses and model experiments

a. Ocean observations

We examine past changes in ocean heat content using new ocean analyses developed at the Met Office (Smith and Murphy 2007). These analyses exploit model-derived covariances to interpolate between observations that are sometimes sparse in space and/or time. Sparsity of ocean observations can limit their use in understanding the variability of the ocean. For comparison purposes, we
also briefly examine analyses from the World Ocean Database (Levitus et al. 2005) and the European Centre for Medium-Range Weather Forecasts (ECMWF) Ocean Reanalysis System 3 (ORA-S3) (Balmaseda et al. 2008). The three ocean analyses make use of similar data but use different methods to analyze the data; thus, differences between them give a crude estimate of the analysis uncertainty. All data are expressed as anomalies relative to a 1961–90 climatological period. Errors in the fall rate of expendable bathythermograph (XBTs; Gouretski and Koltermann 2007; Wijffels et al. 2008) have been corrected in both the Met Office and ECMWF ocean analyses but have negligible impact on North Atlantic temperature anomalies analyzed here (not shown). Note that the depth-independent XBT temperature bias (Gouretski and Reseghetti 2010) has not been corrected.

The spatial patterns of anomalies in ocean analyses could be sensitive to the methodology used to spread the data (i.e., the interpolation method, the choice of covariances, or the use of a dynamical model). Therefore, to give further confidence in the spatial patterns of the analyzed anomalies, we also examine the pattern of anomalies seen in the raw observations: that is, where there has been no infilling of data either spatially or temporally. The quality controlled temperature profiles of expendable bathythermograph (XBTs; Gouretski and Koltermann 2007; Wijffels et al. 2008) have been corrected in both the Met Office and ECMWF ocean analyses but have negligible impact on North Atlantic temperature anomalies analyzed here (not shown). Note that the depth-independent XBT temperature bias (Gouretski and Reseghetti 2010) has not been corrected.

The model used for the experiments for this paper is the Nansen Center version of the Miami Isopycnic Coordinate Ocean Model (MICOM). The model is based on the default MICOM code as described by Bleck and Smith (1990) and Bleck et al. (1992), but it is estimated that two-thirds of the code has been rewritten (Lohmann et al. 2009a). In particular, the Nansen Center version of MICOM conserved heat and salt by the introduction of incremental remapping (Dukowicz and Baumgardner 2000), which ensures the layer thicknesses and tracers are treated in a fully consistent way. The model has been used in several previous studies (for an overview, see Lohmann et al. 2009a), demonstrating skill in reproducing several of the key ocean circulation features in the North Atlantic Ocean.

In this study we use a global configuration at a resolution of approximately 2.4° latitude by 2.4° longitude, with the resolution reduced to 0.8° near the equator [the same model used in Lohmann et al. (2009a,b)]. To avoid a grid singularity in the Arctic Ocean, the North Pole of the grid is located in central Siberia (see Fig. 1b in Furevik et al. 2003). Therefore, the horizontal grid spacing in the North Atlantic is approximately 150 km. In the vertical, 35 isopycnic layers are used, with potential densities ranging from \( \sigma_0 = 21.22 \) to \( \sigma_0 = 28.70 \) and a bulk mixed layer used at the surface.

The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al. 1996) provide radiative fluxes and turbulent surface fluxes of momentum and heat. The forcing scheme and procedure proposed in Bentsen and Drange (2000) is used here. The scheme, which is briefly described in the appendix, reproduces the reanalysis fluxes if the model has the same surface state as in the reanalysis. These states will generally differ and then the fluxes are modified. The turbulent fluxes are modified consistent with the bulk parameterization of Fairall et al. (1996), whereas the longwave radiative fluxes are modified consistent with the Berland and Berland (1952) parameterization. Precipitation is provided by the NCEP–NCAR reanalysis directly, and the evaporation is calculated based on the (modified) latent heat flux. Continental runoff is included by adding freshwater into the appropriate coastal grid cells (Furevik et al. 2003).

The model is spun up from Levitus climatology (Levitus and Boyer 1994; Levitus et al. 1994) by using five consecutive cycles of daily NCEP–NCAR reanalysis fields from 1948 to 2006, yielding a total integration of \(~300\) yr. A sea surface salinity (SSS) relaxation is needed to maintain a stable AMOC. In the spinup, a Newtonian relaxation of SSS toward climatology (Levitus et al. 1994) was applied with an e-folding relaxation time of 30 days for a 50-m-thick mixed layer, which is increased linearly for thicker mixed layers (Bentsen et al. 2004). No relaxation was applied in waters where sea ice is present in March in the Arctic and in September in the Antarctic to avoid spurious relaxations. Also, to avoid extreme fluxes in some regions due to potentially large errors in the models SSS fields, the mismatch between the model and the climatology is limited to \(<0.5\) psu in the calculation of relaxation terms. Finally, a freshwater weekly flux

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1 Also with errors in the XBT fall rate corrected using Wijffels et al. (2008).
correction was computed using the final cycle of spinup. This freshwater flux correction is applied in all the model experiments, with a weak (360 days) SSS relaxation, instead of the 30-day relaxation. No sea surface temperature relaxation is applied.

After the spinup, the model is forced with NCEP–NCAR reanalysis from 1948 to 2006. This experiment will be referred to as the CONTROL experiment. Figure 1 shows the model’s mean Atlantic circulation, calculated by averaging the years 1961–90 from the CONTROL integration. Examining the model’s horizontal circulation (i.e., Fig. 1a), the SPG is a cyclonic westerly intensified gyre, with a peak streamfunction just south of Greenland of $-30$ Sv ($1$ Sv $= 10^6$ m$^3$ s$^{-1}$). The barotropic streamfunction in the CONTROL experiment, as well as in other low-resolution models, is weak compared higher-resolution integrations of the SPG (Treguier et al. 2005; Zhu and Demirov 2011).

The variability in the North Atlantic over the last part of the twentieth century is likely to be complex; a large array of mechanisms is likely to be taking place. To separate the relative importance of different mechanisms, two further sensitivity experiments are also performed. In the first sensitivity experiment, the buoyancy forcing is kept at the seasonally varying climatology, but the historical time-dependent wind stress is used (note that the wind speed used in the bulk formula is also kept at the seasonally varying climatology). This experiment will be referred to as the WIND experiment. In the second sensitivity experiment, the wind stress forcing is kept at its seasonally varying climatology, but the historical time-dependent buoyancy forcing is used (note that the historical time-dependent wind speed is used in the bulk formula). This experiment will be referred to as the BUOY experiment.

Although the effect of only applying one component of the time varying forcing at a time allows us examine the role of different mechanisms, caution is needed in the interpretation of the results. Because of the modification of the surface fluxes by the model (Bentsen and Drange 2000), the WIND experiment does have some variability in buoyancy forcing, especially where sea ice is located (not shown). However, the variability in the buoyancy forcing in the WIND experiment is vastly smaller and considerably damped compared to the BUOY experiment.

3. Evolution of the subpolar gyre in ocean analyses and the CONTROL experiment

Figure 2 illustrates the pattern of recent ocean warming, contrasting the decade 1996–2005 with the previous decade, 1986–95. Between these two decades, a dramatic change is seen in the North Atlantic 0–500-m heat content and SST. The change is particularly dramatic in the region of the SPG, where large negative temperature anomalies in the earlier decade are replaced by large
positive anomalies in the more recent decade. The decadal-mean anomalies calculated from the Met Office ocean analyses compare well with the raw EN3 observations, giving confidence to the pattern of the warming.

Now considering the SPG alone, Fig. 3 shows the evolution of the SPG (50°–66°N, 60°–10°W; see Fig. 1a) upper-ocean heat content in the Met Office, ECMWF, and Levitus ocean analyses. The magnitude of the SPG heat content anomalies is sensitive to the dimensions of the box over which the analysis is averaged, but the temporal variability is not sensitive to varying the boxes limits (not shown). Also shown in Fig. 3 is the inverted wintertime [December–March (DJFM)] NAO index (Hurrell 1995). All indices are normalized to unit variance to aid comparison. Up until 1995, temperatures in the upper SPG are anticorrelated with the NAO index, with maximum correlations (of −0.69, −0.66, and −0.74 for the different analyses) found when the NAO index leads by one year. As the NAO became more positive through the 1970s, 1980s, and 1990s, the SPG cooled as expected given the correlation of NAO and surface heat fluxes (Visbeck et al. 2003).

In the winter (not shown) of 1995/96 the SPG warmed rapidly after a period of intense persistent positive NAO forcing from 1988/89 to 1994/95. The warming coincides with an unusually low NAO index. Following this rapid warming event, there is a persistent offset between the SPG heat content and the inverted NAO index. While the NAO index returns to near neutral levels, the positive heat content anomalies persist, suggesting that some other mechanism influenced the relationship between the NAO and heat content. The heat content is not simply responding to the concurrent or previous year’s NAO conditions.

To better understand the nature of the North Atlantic warming in the 1990s, we examine upper-500-m heat content anomalies (Fig. 4). Figure 4 (left) shows the evolution in four pentads as seen in the Met Office analyses. From 1985 to 1989, the SPG is anomalously cold (relative to the 1961–90 reference period), and there is a slight warm anomaly in the northwest subtropical gyre. By 1991–95, corresponding to the highest NAO values (see Fig. 3), both these anomalies have intensified greatly, and the warm anomaly in the northwest subtropical gyre has begun to spread eastward. In 1996–2000, the rapid warming of the SPG is evidenced by the disappearance of the cold anomalies seen in the previous pentad, and the warm anomaly is reaching northward into the eastern SPG. Finally, in 2000–04, as the overall warming of the SPG continues, this warm anomaly spreads round the eastern flank of SPG, reaching into the Irminger Sea.

The temporal evolution of the heat content in the analyses is largely reproduced in the raw EN3 observations.
especially the evolution of warm anomalies from the northern subtropical gyre into the SPG. However, there is some disagreement between the analyses and the raw EN3 observations in the fine structure of the anomalies and also the magnitude of some anomalies. Minor differences would be expected because of the spreading of data in the analyses and the limited number of data points in the average of raw observations. Overall, the comparison suggests that the spatial pattern of anomalies associated with the warming is robust.

The evolution of upper-500-m heat content anomalies in the CONTROL experiment is very similar to that seen in the analyses (see Figs. 4e–h). However, there is some disagreement between the analyses and the raw EN3 observations in the fine structure of the anomalies and also the magnitude of some anomalies. Minor differences would be expected because of the spreading of data in the analyses and the limited number of data points in the average of raw observations. Overall, the comparison suggests that the spatial pattern of anomalies associated with the warming is robust.

The agreement between the evolution of ocean heat content anomalies at the surface and at depth between the observations and the model experiments suggests that similar mechanisms are responsible. The next section will explore the changes in the model in order to understand the key mechanisms in greater detail.

4. What are the mechanisms responsible for the warming?

a. Evolution of heat content anomalies in the North Atlantic

FIG. 3. Flipped (i.e., multiplied by −1) normalized DJFM NAO index based upon the difference of normalized SLP station data between Lisbon and Reykjavik (black; NAO index data provided by the Climate Analysis Section, NCAR, Boulder, Colorado; Hurrell 1995). Normalized 0-500-m annual average temperature anomalies (°C), relative to the mean for 1961–90, for the SPG (50°–66°N, 60°–10°W) from Met Office (red), ECMWF (blue), and Levitus (green) ocean analyses. One standard deviation corresponds to 2.1 for the NAO index and 0.256, 0.253, and 0.251 for the Met Office, ECMWF, and Levitus SPG indices, respectively. Note that the temperature of the SPG and the NAO are anticorrelated to 1995. After 1995, the correlation breaks down, with the warming seen to be more persistent than a simple correlation with the NAO would suggest.

As the NAO reached its peak values in the early 1990s (Fig. 3), a large layer of anomalously cold Labrador sea-water was formed (Yashayaev et al. 2007) because of increased deep convection in the Labrador sea, particularly in the years 1990–93 (Lazier et al. 2002). Figure 5 shows the 1000–2500-m temperature anomalies from the observed datasets. In the observations, there was a significant cooling of the deep SPG between 1986 and 1990 and between 1996 and 2000, with some suggestion that cool temperature anomalies have propagated southward along the western boundary (see Figs. 5a–d). Increased deep convection is also seen in the CONTROL experiment at this time, although largely in the Irminger Sea, and the deep temperature anomalies show similar evolution to the observed (see Figs. 5e,f). Initially, large temperature anomalies form in the Labrador and Irminger Basins after the onset of the positive NAO, before intensifying in the early 1990s and propagating southward along the western boundary.

The agreement between the evolution of ocean heat content anomalies at the surface and at depth between the observations and the model experiments suggests that similar mechanisms are responsible. The next section will explore the changes in the model in order to understand the key mechanisms in greater detail.

2 Note the model anomalies are for a 1500–2500-m layer because the warm surface anomalies penetrate deeper in the model than in the observed datasets.
FIG. 4. Comparison of observed and modeled 0-500-m ocean average temperature anomalies (°C) relative to the mean for 1961–90. (a)–(d) Pentadal-mean anomalies from 1986 to 2005 from the Met Office ocean analysis. (e)–(h) Pentadal-mean anomalies calculated only from the EN3 profile data. (i)–(l) As in (e)–(h), but for the CONTROL experiment.
downwelling Rossby waves (Schneider et al. 2002; Leadbetter et al. 2007), and also in the eastern subpolar gyre from 1996 onward (Figs. 6g,h), which could be consistent with changes in the SPG boundaries due to wind stress curl anomalies (Häkkinen et al. 2011). However, the warm anomalies are smaller in magnitude than those that are seen in the CONTROL experiment. Thus, the wind forcing of Rossby waves, as well as of the gyre extent, was not responsible for the major features seen in the evolution of temperature anomalies in the North Atlantic.

**Figure 6** shows that the buoyancy forcing is key to the development of the spatial patterns of observed heat content anomalies in the North Atlantic between 1986 and 2005. But is the ocean heat content simply reacting to changes in local heat fluxes? Closer examination of the surface heat flux anomalies given to the ocean in the BUOY experiment, relative to the 1961–90 mean calculated from the CONTROL experiment, shows that the atmosphere is not the cause of the spatial pattern of temperature anomalies. The surface heat fluxes are often acting to damp the positive heat content anomalies in the North Atlantic Current in the late 1980s and early 1990s (see Fig. 7). The damping is especially clear in the eastern SPG following the warming in the mid-1990s (see Figs. 7c,d). Thus, the spatial pattern of heat content anomalies seen in Fig. 4 is not created by the surface heat fluxes and must result from a dynamical change in the ocean circulation forced by anomalous buoyancy fluxes.

### b. Subpolar gyre heat content variability

Examining the evolution of 0–500-m heat content anomalies averaged over the SPG (i.e., 50°–66°N, 60°–10°W; see Fig. 8a), we see that the CONTROL experiment largely captures the sign and magnitude of the temporal variability from 1960 to 2005. The timing of the rapid warming of the SPG in the winter of 1995/96 is also captured but is larger than that seen in the observations. Although some error might be expected in the CONTROL experiment simulation, because of the imperfect nature of the model, it is also worth considering the effect of the ocean analyses on the observed warming. The magnitude of the anomalies associated with the warming in the raw EN3 observations is somewhat greater than that seen in the Met Office analysis (Fig. 4) and could therefore point to a failing of the analysis procedure leading to smaller anomalies in the ocean analyses.

The 0–500-m SPG heat contents in the BUOY and WIND experiments show some interesting differences in their evolution when compared to the CONTROL experiment. Figure 8a shows that the cooling of the SPG in

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**Fig. 5.** Comparison of observed and modeled 1000–2500-m average temperature anomalies (°C) relative to the mean for 1961–90. The 1986–90 mean anomalies for (a) the Met Office ocean analysis, (b) the raw EN3 profiles, and (c) the CONTROL experiment. (d)–(f) As in (a)–(c), but for the 1996–2000 mean. Note that temperature anomalies are averaged from 1500 to 2500 m for the CONTROL experiment (see text for details).
the early 1980s is primarily driven by the changes in the buoyancy forcing. However, the changes in the wind stress also seem to play a role in the cooling as the positive NAO reaches its peak in the late 1980s and early 1990s (this will be discussed later). Both the buoy and WIND experiments show rapid increase in heat content around 1995/96. However, as expected from Fig. 6, the magnitude of the warming seen in the CONTROL experiment is only recreated in the BUOY experiment. The warming seen in the WIND experiment is associated with the development of weak positive heat content anomalies in the eastern SPG in 1996 and 1997 (Figs. 6g,h).

To illustrate the importance of the change in the ocean circulation, compared to the local surface heat fluxes, the heat budget of the North Atlantic SPG is shown in Fig. 9. For the purpose of this calculation, it is assumed that the heat budget is defined as

\[ \Delta E = H_O - H_A, \]

where \( \Delta E \) is the change in annual-mean temperature (expressed as energy) integrated over a box covering the SPG (from the surface to the bottom of the ocean) calculated by subtracting the previous annual-mean temperature from the current year. The terms \( H_O \) and \( H_A \) are the oceanic and atmospheric fluxes, respectively, and are calculated by summing the monthly fluxes of energy between the two annual means used to calculate \( \Delta E \). The term \( H_O \) is the energy delivered to the SPG box due to convergence of oceanic heat transport between the two annual means and is defined as the energy transported into the southerly edge of the box, minus the energy transported out of the box at its most northerly extent. Finally, \( H_A \) is the energy removed from the box by the atmosphere due to the surface heat flux integrated over the SPG.

Figure 9a shows the temporal evolution of \( H_O \) and \( H_A \) in the model experiments. The difference between these two terms gives the change in energy of the SPG, which is shown in Fig. 9b (solid lines). To assess that the heat budget is closed, also plotted in Fig. 9b is the \( \Delta E \) computed directly from the model’s temperature field (dashed lines). The comparison shows that the budget is not perfectly closed, which is probably due to the use of monthly means to calculate the ocean heat transports. Nevertheless, the \( \Delta E \) that is calculated from \( H_O \) and \( H_A \) is a good fit to the modeled changes.

3 Note that, in order to simplify the calculation of the heat budget, the SPG region for the calculation is changed slightly. The SPG is now defined as the integrated volume across the whole Atlantic in the latitude band 50°–65°N (i.e., 50°–65°N, 60°W–20°E). The evolution of SPG heat content is not changed substantially by integrating across the basin (not shown).
Throughout most of the CONTROL experiment (see Fig. 9a, black lines), the amount of energy that is removed from the ocean column by the atmosphere has similar magnitude to the convergence of the ocean heat transport. Therefore, deviations in the column-integrated temperature are usually modest (Fig. 9b). The largest deviations away from a balanced heat budget occur in the mid-1990s. The difference in the calculated \( H_O \) and \( H_A \) is \( \sim 1 \times 10^{22} \text{ J} \) in 1996 (Fig. 9b). Although both \( H_O \) and \( H_A \) contribute to the warming, the largest contributor to the change in energy is a large surge in the oceanic heat transport convergence \( H_O \) from \( \sim 0.9 \times 10^{22} \text{ J yr}^{-1} \) in 1990 to a peak of \( 1.75 \times 10^{22} \text{ J yr}^{-1} \) in 1996. Coincident with the surge in \( H_O \) is a decrease in \( H_A \). The \( H_A \) is anomalously high in 1993/94 at \( \sim 1.1 \times 10^{22} \text{ J yr}^{-1} \) but drops to \( \sim 0.7 \times 10^{22} \text{ J yr}^{-1} \) in 1996/97. The drop in \( H_A \) is \( \sim 40\% \) of the change in \( H_O \), but the change is exaggerated by the large values of \( H_A \) in the early 1990s. There is nothing unusual in the values of \( H_A \) at the time of the warming, but the change in \( H_O \) is highly unusual. Following the rapid warming, \( H_O \) remains larger than \( H_A \) in the early 2000s, making \( \Delta E \) positive (Fig. 9b).

Figure 9a shows that changes in the wind stress in 1996/97 contribute to the surge in \( H_O \). However, the increase in \( H_O \) in the WIND experiment is only a fraction of that seen in the CONTROL experiment, leading to only a peak \( \Delta E \) of \( \sim 0.4 \times 10^{22} \text{ J yr}^{-1} \) in 1995/96, which quickly decreases (Fig. 9b). In comparison, a larger and more sustained increase in \( H_O \) is captured in the BUOY experiment, which leads to a peak \( \Delta E \) of \( \sim 0.6 \times 10^{22} \text{ J yr}^{-1} \). Importantly, the time-integrated difference between the BUOY and WIND experiments is much larger than the difference in the peak values. Following the year 2000, the BUOY experiment still has a positive \( \Delta E \) of \( \sim 0.25 \times 10^{22} \text{ J yr}^{-1} \) because of the increased \( H_O \).

**Fig. 7.** The 5-yr-mean surface heat flux anomalies from the BUOY experiment (W m\(^{-2}\)), relative to the 1961–90 mean computed directly from the CONTROL experiment. Negative values indicate a net flux out of the ocean.
c. Changes in the ocean circulation and heat transport

In this section, we examine the changes in ocean heat transport further and their relationship to changes in the gyre and overturning circulations. Previous studies have suggested that the SPG transport increased into the mid-1990s (Häkkinen and Rhines 2004; Hátún et al. 2005). Figure 8b shows the evolution of the SPG barotropic streamfunction anomaly (calculated by averaging the anomalies over the SPG, 50°–66°N, 60°–10°W; see Fig. 1a). The SPG barotropic transport in the CONTROL experiment strengthens from the 1980s up until the mid-1990s, to a peak of ~3 Sv immediately before the warming. At the time of the warming the SPG barotropic transport quickly declines, similarly to the observations (Häkkinen and Rhines 2004). The decline in the SPG barotropic streamfunction is associated with an
anticyclonic anomaly that appears in the east SPG in 1995/96 (not shown). The anticyclonic anomaly is located where warm anomalies propagate northward into the SPG (not shown) and is related to a contraction of the SPG consistent with Hátún et al. (2005).

There are also changes in the overturning circulation in the North Atlantic. Figure 8c shows the changes in the AMOC at 45°N in the experiments. The AMOC in the CONTROL experiment shows a strengthening trend in the early 1990s, before weakening (see Fig. 8c). Similar changes in the high latitude AMOC are seen in other ocean reanalyses (Anderson et al. 2009; H. Pohlmann et al. 2012, unpublished manuscript) and similar ocean-only experiments (Böning et al. 2006; Deshayes and Frankignoul 2008; Grist et al. 2010).

Closely associated with the increase in the AMOC in the CONTROL experiment is an increase in the oceanic meridional heat transport (MHT) in the North Atlantic (Fig. 10d), although the correlations are not perfect (the mechanisms of the heat transport changes will be investigated further later on). The MHT at the subpolar–subtropical gyre boundary increases in the late 1980s to a peak in the mid-1990s at $-2 \times 10^{14}$ W but decreases at the northern edge by $-0.5 \times 10^{14}$ W. Therefore, the increased heat transport convergence in the SPG in the mid-1990s (see Fig. 9) is largely driven by the increase of the MHT at the southern boundary of the SPG. Examining the heat transport at the southern edge of the SPG in more detail, the peak MHT anomaly of $\sim 3.5 \times 10^{14}$ W in the CONTROL experiment is found to be associated with a $\sim 1.25 \times 10^{14}$ W anomaly in the upper, northward-flowing limb of the AMOC and a $\sim 1.25 \times 10^{14}$ W anomaly in lower, southward-flowing limb (not shown). This vertical structure in the heat transport anomalies is consistent with a key role for the AMOC.

The sensitivity experiments clearly show that the changes in the ocean circulation are driven by the changes in the buoyancy fluxes, and not the wind stress. The BUOY experiment captures the CONTROL experiment changes in the SPG circulation (Fig. 8b), Greenland–Scotland Ridge overflow transport (not shown), and the strengthening of the AMOC (Fig. 10), but the WIND experiment does not. Examining the MHT in the sensitivity experiments, it is clear that the major changes in the MHT are due to changes in the AMOC. The change in the heat transport in the CONTROL experiment is only recreated in the BUOY experiment: that is, when there is a large-scale change in the ocean circulation (see Fig. 9d). Similarly to the AMOC variability, the MHT variability in the WIND experiment is much higher frequency than in the BUOY experiment (see Fig. 9f). Importantly, there is no sustained increase in the MHT in the subpolar gyre region in the WIND experiment.

d. Partitioning the ocean heat transport

To investigate the changes further, the ocean heat transport is broken down into components that are associated with the mean circulation and temperature field and anomalies associated with changes in the strength of the ocean circulation, mean advection of temperature anomalies, and the covariance of the anomalies in

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4 Calculated as the integral over depth and longitude of velocity and temperature at each latitude.
circulation and the temperature. Integrating the velocity $v$ and temperature $T$ across the Atlantic gives the total heat transport, which as in Dong and Sutton (2002) can thus be broken down as

$$\rho c_p \int_0^H \int_W v T \cos \delta \delta z = \rho c_p \int_0^H \int_W \bar{v} T \cos \delta \delta z + \rho c_p \int_0^H \int_W v' T \cos \delta \delta z,$$

$$+ \rho c_p \int_0^H \int_W \bar{v} T' \cos \delta \delta z + \rho c_p \int_0^H \int_W v' T' \cos \delta \delta z.$$
where $\nu$ is the full meridional velocity, $T$ is the temperature, $\bar{T}$ or $\bar{\nu}$ represent the climatological values at each grid point, and the $'$ is the anomaly relative to the climatology. Here, the climatology period is 1961–90. The term $\rho$ is a constant reference density, $H$ is the depth of the ocean, and $c_p$ is the specific heat capacity at constant pressure. The heat transports are calculated from the monthly-mean $\nu$ and $T$ data.

Figure 11 shows the anomalous components of MHT associated with $\nu'$ and $T'$ at $40^\circ$ and $50^\circ$N in all three experiments. The heat transport in the CONTROL experiment is largely flat between 1960 and 1980 at both $40^\circ$ and $50^\circ$N but increases from the early to mid-1980s. At $40^\circ$N, the MHT variability is dominated by changes in the circulation ($\nu'T'$). However, at $50^\circ$N the components associated with the advection of temperature anomalies (i.e., $\nu'T'$) become important.

At $50^\circ$N, the total heat transport associated with all components related to $\nu'$ and $T'$ increases suddenly from $\sim1 \times 10^{14}$ W in 1994 to $\sim2.5 \times 10^{14}$ W in 1995/96 (Fig. 11a). The surge in the MHT at $50^\circ$N in the CONTROL experiment is largely associated with a sudden increase in the strength the circulation (i.e., $\nu'T$), which increases to $\sim1.5 \times 10^{14}$ W, approximately 60% of the total MHT anomaly. The increase in the $\nu'T$ is not due to the strengthening of the SPG but occurs after the peak in the barotropic streamfunction (Fig. 8b).

After the rapid warming of the SPG, the MHT component that is associated with covarying changes in the ocean circulation and temperature (i.e., $\nu'T'$) becomes important at $50^\circ$N. This component is negligible at most latitudes and for most of the integration (not shown). The increase in $\nu'T'$ is related to changes in the SPG gyre extent and spreading of subtropical waters northward. Thus, changes in the gyre extent did not cause the surge in ocean heat transport in 1995/96. However, the increase in $\nu'T'$ to a peak in the early 2000s does suggest that the change in gyre extent was important for the persistence of the warming of the SPG after the mid-1990s.

The increase in the MHT seen in the CONTROL experiment is captured largely by the BUOY experiment but not in the WIND experiment. Similarly to the CONTROL experiment, the BUOY experiment sees a large increase in the MHT at $40^\circ$N (Fig. 11e), largely due to $\nu'T$, and also at $50^\circ$N, where $T'$ becomes important (Fig. 11b). In contrast, the WIND experiment sees no significant low-frequency increase in the MHT (Figs. 11c,f), and negative MHT anomalies are present at $50^\circ$N before the mid-1990s. The negative MHT in the WIND experiment suggests that the wind stress partially offset
the increase in the buoyancy-forced MHT, partly ensuring that the SPG remained cold. A proportion of the surge in the MHT associated with $\nu T$ at 50°N in the CONTROL experiment in 1995/96 is due to a short-lived contribution from changes in wind stress (Fig. 11c). However, the surge in $\nu T$ in the WIND experiment has a similar magnitude to the contribution from the buoyancy forcing (Fig. 11b) at this time.

5. Discussion

The analysis of the previous section has shown that the pattern of warming of the North Atlantic SPG in the mid-1990s was primarily caused by changes in the ocean circulation that were forced by the buoyancy fluxes, not changes in the wind stress. The buoyancy fluxes associated with the positive NAO forced a strengthening of the Atlantic circulation and in particular the AMOC in the 1980s into the 1990s. The change in ocean circulation led to an increase in the ocean heat transport, which was key to the warming of the SPG. Therefore, the warming of the SPG was largely a consequence of the prolonged positive NAO forcing that preceded it (i.e., hypothesis 2). This result expands on the work of Lohmann et al. (2009b), who suggested that the SPG would have weakened from 1995 in absence of the negative NAO.

Undoubtedly, the decline in the NAO after the mid-1990s also had a role in the warming. In particular, the strong negative NAO in 1995/96 forced a temporary reduction in the surface heat flux from the SPG in 1996 and a temporary increase in the MHT, probably due primarily to a sudden relaxation of southerly Ekman currents (see Fig. 11c). However, the surge in MHT due to the negative NAO (i.e., in 1995/96) was not as large as the increase in the MHT due to the preceding buoyancy forcing associated with the positive NAO.

Although it is clear that the negative NAO of 1995/96 impacted on the rapidity and magnitude of the SPG warming, the experiments suggest that the preceding positive NAO was key. Idealized experiments also suggest that the warming was inevitable before the negative NAO took place. Forcing the ocean model with persistent positive NAO also forces a response that is similar to the observed changes, but the warming of the SPG is not as rapid (Lohmann et al. 2009a; Robson 2010). Therefore, the results strongly suggest that without the strengthening of the AMOC the warming of the SPG following the negative NAO of 1995/96 would have been much smaller and shorter in duration.

The conclusion that the buoyancy forcing, not the wind stress, was the most important driver of the changes seen in the subpolar North Atlantic is in disagreement with some other recent studies, in particular Lorbacher et al. (2010) and Häkkinen et al. (2011), which suggest that the wind stress changes may have played a more important role. However, the response to buoyancy forcing is not considered in either Lorbacher et al. (2010) or Häkkinen et al. (2011). Lorbacher et al. (2010) argue that the wind forcing due to the negative NAO in 1995/96 is important for driving changes on the interannual time scale over the whole North Atlantic. However, close examination of their Fig. 4 shows that the wind-forced experiment does not recover the linear 1993–99 sea surface height trend in the SPG region. Examining our own experiments further, we see that the observed spatial pattern of the 1993–99 SPG sea surface height trend shown in Lorbacher et al. (2010) (see their Fig. 4a) is also largely recovered in the CONTROL and BUOY experiments (not shown). The changes are due to both changes in the wind and buoyancy forcing, but the changes in the SPG, especially in the eastern SPG, are primarily due to the changes in the buoyancy forcing.

As with any modeling study, there are caveats to this study. Variability in the wind stress or buoyancy forcing may project differently onto the model’s circulation due to errors in the mean circulation or deep-water formation sites. Also, the model does not resolve ocean eddies, which may be important for the interannual variability of the ocean heat transport (Volkov et al. 2008) and the AMOC (Biastoch et al. 2008; Kanzow et al. 2009). The speed at which AMOC anomalies propagate is also model dependent (Getzlaff et al. 2005; Hodson and Sutton 2012). The initial adjustment of the ocean is usually faster in higher-resolution models (Döösher et al. 1994; Getzlaff et al. 2005; Roussenov et al. 2008; Hodson and Sutton 2012), presumably because of an improved simulation of boundary waves (Döösher et al. 1994); in particular, the Kelvin wave speed is sensitive to model resolution (Hsieh et al. 1983) and coastally trapped waves are sensitive to the local stratification (Wang and Mooers 1976). Given the low resolution of the model, we might therefore expect the southward propagation of AMOC anomalies to be slow compared to higher-resolution models. However recent results from Zhang et al. (2011) show that high-resolution models can also support slow propagation of AMOC anomalies.

Finally, the fact that we have used an ocean model rather than a coupled model has inherent limitations. We have not been able to study any coupled ocean–atmosphere aspects of the evolution. The need for prescribed freshwater fluxes and weak SSS relaxation to maintain the mean strength of the overturning circulation is also a limitation of this study. The weak SSS relaxation could also impact on the AMOC variability at decadal time scales and modulate the model’s response to the NAO forcing.
6. Summary and conclusions

This paper has investigated the causes of the rapid warming of the North Atlantic in the mid-1990s, especially the sudden warming of the North Atlantic subpolar gyre (SPG) following the 1995/96 negative NAO index. The event was explored by a comparison of ocean observations with model experiments to identify the important mechanisms involved. The main results are as follows:

- The rapid warming of the North Atlantic SPG was primarily a result of a surge in northward ocean heat transport in the mid-1990s.
- The key changes in the ocean heat transport were primarily caused by changes in the ocean dynamics in response to the surface buoyancy forcing of the SPG over the previous decade. More specifically, a strengthening of the Atlantic circulation, especially the Atlantic meridional overturning circulation (AMOC), following the persistent positive NAO index in the late 1980s and early 1990s was crucial for the observed changes.
- The wind stress forcing had a much smaller effect on the development of heat content anomalies than the buoyancy forcing changes. The wind stress did act to modulate the changes in the ocean heat transport, especially during the negative NAO in 1995/96, which contributed a short-lived increase in the ocean heat transport. Consequently, the wind forcing contributed to the rapidity of the warming of the SPG but was not the primary cause.
- A short-lived reduction in the surface heat flux over the SPG in 1995/96, associated with the low NAO index, also contributed to the warming. However, these changes were much less important that the ocean heat transport.
- At latitudes around 40°N, the variation in ocean heat transport is dominated by increases in the mean circulation of the model \((\nu T)\). However, at 50°N the components of the ocean heat transport associated with temperature anomalies become important, partly associated with the contraction of the SPG after 1995/96 (i.e., \(\tau T\) and \(\nu T\)).

We present persuasive evidence that the AMOC has played a central role in the recent decadal variability of the North Atlantic, including its ability to bring about a “rapid change” event. Because the warming was largely a lagged response to the positive NAO, there are important implications for decadal predictability. The North Atlantic SPG has already been highlighted as a region where initialized climate predictions show improved skill relative to predictions that do not assimilate data at predicting ocean heat content anomalies (Smith et al. 2010; Robson 2010). However, it is important to understand the origin of the improved skill. Robson 2010 shows that the Met Office’s Decadal Prediction System (DePreSys) is able to predict the rapid warming of the SPG in the mid-1990s. Analysis is ongoing, but the initialization of a strong AMOC is important for the skillful predictions.

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APPENDIX

Surface Fluxes

The turbulent fluxes for momentum \(\tau\), sensible heat \(Q_H\), and latent heat \(Q_L\), used by the model are calculated by the scheme of Bentsen and Drange (2000). The calculation takes place in two stages. First, an iterative scheme estimates the atmospheric state at a height of 10 m. More specifically, the turbulent fluxes and surface data (sea surface temperature and specific humidity) from NCEP–NCAR reanalysis are inverted using the bulk expressions of the Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) scheme (Fairall et al. 1996) to give an estimate of the atmospheric state for each time step (i.e., \(S_o\), \(u_o\), \(\theta_o\), and \(q_a\)). Here \(S_o\) is the average wind speed, \(u_o\) is the mean wind vector, \(\theta_o\) is the potential temperature, and \(q_a\) is the specific humidity. The \(\theta_o\), \(q_a\), and the surface pressure from NCEP–NCAR reanalysis are then used to update the air density \(\rho_a\), using an equation of state for moist air (Gill 1982).

Once the atmospheric state has been estimated to the desired accuracy, the turbulent fluxes that are given to the model are then calculated. The calculation is again performed using the TOGA COARE scheme, but now using the estimated atmospheric state and the surface data from the ocean model. Therefore, if the models surface state is the same as that in NCEP, the scheme should reproduce the same fluxes.
The turbulent fluxes used by the model are given as

$$
\tau = -\rho_a C_{D,m} S_a u_a \\
Q_H = \rho a C_{H,m} S_a (T_{s,m} - \theta_a), \quad \text{and} \\
Q_L = \rho a L_c C_{E,m} S_a (q_{s,m} - q_a),
$$

where $C_{D,m}$, $C_{H,m}$, and $C_{E,m}$ are the transfer coefficients for momentum, sensible heat, and latent heat, respectively, calculated for the model. The term $c_a$ is the specific heat capacity of air and $L_c$ is the latent heat of vaporization. Finally, $T_{s,m}$ and $q_{s,m}$ are the sea surface temperature and specific humidity taken from the model.

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