PALEOCEANOGRAPHY, VOL. 22, PA3104, doi:10.1029/2007PA001439, 2007

Impact of extended NAO buoyancy forcing on the subpolar North Atlantic and climate variability over the last millenium

Duo Yang[1] and Paul G. Myers[1]

Received 23 February 2007; revised 16 July 2007; accepted 31 July 2007; published 28 September 2007.

[1] We examine the impact of forcing a regional eddy-permitting ocean model of the subpolar North Atlantic with anomalous buoyancy fluxes corresponding to extended extreme states of the North Atlantic Oscillation (NAO). We find a weakened (enhanced) Meridional Overturning Circulation (MOC) in our low (high) NAO simulations. Such results may be consistent with ideas that suggest the Medieval Warm Period (MWP) and the Little Ice Age are associated with periods when the NAO is predominantly in one phase. Our results also show multidecadal MOC variability with a period of 70 years in our high-NAO simulation, in comparison to an equilibrium-like circulation under low-NAO related forcing. A less stable climatic state is therefore suggested to occur under high-NAO forcing, which may be consistent with ideas that suggest the warming associated with the MWP did not occur synchronously over the North Atlantic. The variability appears to be brought about by internal oceanic processes.

Citation: Yang, D., and P. G. Myers (2007), Impact of extended NAO buoyancy forcing on the subpolar North Atlantic and climate variability over the last millenium, *Paleoceanography*, *22*, PA3104, doi:10.1029/2007PA001439.

1. Introduction

[2] The North Atlantic Oscillation (NAO) is the most prominent and fundamental mode of atmospheric and climate variability over the North Atlantic Ocean [*Hurrell et al.*, 2003]. *Hurrell* [1995] originally constructed his NAO time series on the basis of data going back to 1864. Others have since extended the instrumental record back to the early nineteenth century [e.g., *Jones et al.*, 2001]. Many authors examining North Atlantic paleoclimatic records varying from tree rings [*Cook et al.*, 1998] to Greenland ice accumulation [*Appenzeller et al.*, 1998] to stalagmite growth [*Proctor et al.*, 2002] have found climatic variability that they linked in part to the NAO, thus suggesting that the NAO index could be extended into the past through proxy records [*Luterbacher et al.*, 2002]. Multiproxy records of the NAO now extend back at least to the middle of this millenium [*Luterbacher et al.*, 2002; *Cook et al.*, 2002].

[3] An interesting feature seen in many of these extended time series is the observation that they remain predominantly in one phase for an extended period of time. Several studies [e.g., *Proctor et al.*, 2000; *Cook et al.*, 2002] have suggested that the NAO was in a predominantly high phase for a number of centuries prior to $\sim$1300, before switching to an extended period of low values in the following centuries [e.g., *Proctor et al.*, 2000; *Luterbacher et al.*, 2002; *Clarke and Rendell*, 2006]. *Bruckner and Mackensen* [2006] discuss the speculation that this is linked to historically known climatic periods, such as the Medieval Warm Period (MWP) and the Little Ice Age (LIA).

[4] *Mann* [2003] points out that the evidence for the MWP as a period of global (or even hemispheric) warmth is weak. Similar issues may also exist with the LIA [*IPCC*, 2001]. *Mann* [2003] and others instead speculate on a changing oceanic circulation modifying regional North Atlantic climate. *Keigwin and Pickart* [1999] suggest a northward movement of the slope water system north of the Gulf Stream associated with an extended period of low-NAO forcing during the Little Ice Age. Present day observations show that Labrador Sea Water (LSW) formation and the North Atlantic Current (NAC) have displayed significant interannual and interdecadal variability, with changes in convection, thinning and warming of LSW linked to changes in the NAO index [*Curry et al.*, 1998]. There has also been speculation of a link between changes in ocean circulation and the Atlantic Multidecadal Oscillation (AMO) [*Knight et al.*, 2005].

[5] Studies from Ocean General Circulation Models (OGCMs) have linked NAO-related interannual or interdecadal atmospheric forcing to low-frequency variations in the horizontal and meridional circulation and heat transport [e.g., *Eden and Willebrand*, 2001]. However, a series of idealized models forced by idealized, stable flux forcing did exhibit internal oceanic oscillations, independent of varying surface forcing [e.g., *te Raa et al.*, 2004].

[6] We hence inspect this issue in a realistic (in terms of geometry and topography) and eddy-permitting regional ocean model configuration. With better resolved LSW

[1]Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alberta, Canada.

0883-8305/07/2007PA001439

formation and NAC pathway in this resolution of the model (eddy-permitting), we reveal interactions of water mass formation and boundary currents, as well as their role in the initiation and maintenance of an internal oceanic mode. We note the different stability of the ocean during extended periods of high- and low-NAO forcing.

2. Model Configuration and Experiment Design

[7] The model is a flux-forced version [*Yang*, 2005] of a regional eddy-permitting (1/3°) ocean model of the subpolar North Atlantic (68°W $\sim$ 0° and 38 $\sim$ 70°N) [*Myers*, 2002]. The climatological surface buoyancy forcing originates from the Southampton Oceanography Center (SOC) [*Josey et al.*, 1998]. Wind stress is from an European Center for Medium Range Weather Forecast (ECMWF) climatology [*Trenberth et al.*, 1990]. The northern boundary is closed with restoring buffers in Hudson Strait, Baffin Bay and the region to the north of Iceland. At the open southern boundary, the tracers are allowed to be advected or diffused out of the domain if the velocities (perpendicular to the boundary) at some level are directed out of the domain, and are restored smoothly to climatology if the velocities are directed into the domain. The model, which is described in more detail by *Myers* [2002] and *Yang* [2005], is spun up for 40 years from rest (and climatological temperature and salinity). The major features of the large-scale circulation of the subpolar North Atlantic are well represented except our "Northwest Corner" (near 51°N 43°W where the NAC takes a 90° turn to the right, defined originally by *Worthington* [1976]) is a bit far to the east, as shown by *Yang* [2005]. The model basin averaged salinity field also drifts, as is common in these type of model [e.g., *Treguier et al.*, 2005].

[8] Two experiments are run from the end of the spin-up. Named HNAO and LNAO, they are forced by anomalous monthly heat and freshwater fluxes corresponding to extreme high- and low-NAO phases superposed upon the climatological forcing (wind stress unchanged), and were integrated 90 years. HNAO was then integrated for a further 110 years to allow for an analysis of the greater variability discussed in section 3. We emphasize the effect of the buoyancy fluxes (and thus water formation changes) by leaving the wind stress unchanged but note that the impact of NAO related wind stress changes is implicitly included in the flux data used. The extreme NAO phase related anomalous heat and freshwater fluxes are synthesized from three extreme high-NAO index years (1989, 1990, and 1995) in HNAO and from three extreme low-NAO index years (1969, 1996, and 1963) in LNAO, taken from the NCEP/NCAR reanalysis [*Kalnay et al.*, 1996]. The December–February (DJF) heat flux anomalies show an obvious seesaw pattern (Figure 1). Heat loss is most intensive over the Labrador Sea and decreases toward the southeast during the extreme high-NAO phase, whereas it is strongest in southeast corner of our domain, around the Bay of Biscay, and reduces toward the Labrador Sea in the extreme low-NAO phase. The DJF averaged heat flux difference between the two extreme NAO phases over the Labrador Sea reaches 200 $\mathrm{Wm^{-2}}$.

3. Model Response of the Subpolar North Atlantic and Mechanism

3.1. Less "Stable" Oceanic Response in HNAO Versus LNAO

[9] The meridional stream function (MSF) at 55°N (Figure 2a) reveals a strong circulation in HNAO compared to LNAO. Additionally the MSF, the domain and eastern basin (43–52°N 12–24°W) averaged heat content (Figure 2) reveal a multidecadal oscillation in HNAO while a quasi-equilibrium is established in LNAO following an initial adjustment. Interannual variability is ubiquitous in both experiments but the HNAO shows a large and distinct oscillation with a period of around 70 years. The LNAO experiment was terminated after being integrated through the length of the oscillation in HNAO (90 years) without showing any signs of such variability.

[10] The MSF at 55°N mirrors to a great extent the strength and timing of Labrador Sea convection in the model. Figure 2 shows that the changes in the MSF lead variations in the domain (and the eastern basin) heat content by 4 (6) years, suggesting a leading role of Labrador Sea convection. The maximum correlation coefficients between the time series are 0.73 (0.66), significant at the 99% level on the basis of the effective number of degrees of freedom.

[11] Nevertheless, since there is no variability other than the seasonal cycle in our surface forcing, the different evolving patterns, especially the multidecadal variability appearing in HNAO, must be brought about by internal oceanic processes. Figure 3 gives snapshots of annual temperature anomalies in HNAO over a full oscillation (years 46–94). After switching to anomalous high-NAO forcing in year 41, the increased heat loss over the Labrador Sea produces colder and hence denser surface water, enhancing deep vertical convection (reaching 1700 m) and producing a greater volume of LSW. This leads to a greater proportion of the NAC transport being directed toward the Labrador Sea. As a consequence, the heat content drops in the eastern basin while the Labrador Sea warms (e.g., year 46 in Figure 3). The warming of the Labrador Sea leads to a slow reduction in the depth of convection (to a minimum of 1100 m) and decreased LSW production. As convection weakens in the Labrador Sea, the pathway of the NAC shifts more to the east, reducing the transport of warm salty water into the Labrador Sea while increasing this transport into the eastern basin (e.g., during years 52–80, see Figure 3 for sample years 58 and 70). However, the cooling of the Labrador Sea is sufficient that the depth of convection eventually starts to increase again and the oscillation repeats over the next decades (e.g., year 94 in Figure 3 and also see Figure 2c).

[12] The low-frequency variability in HNAO is therefore a function of the changing velocity fields in the model upper layers, leading to anomalous heat transports (the switch of the NAC pathway described above) and regional temperature anomalies. The advected heat then regulates convection in the model, modifying the basin-scale pressure fields,

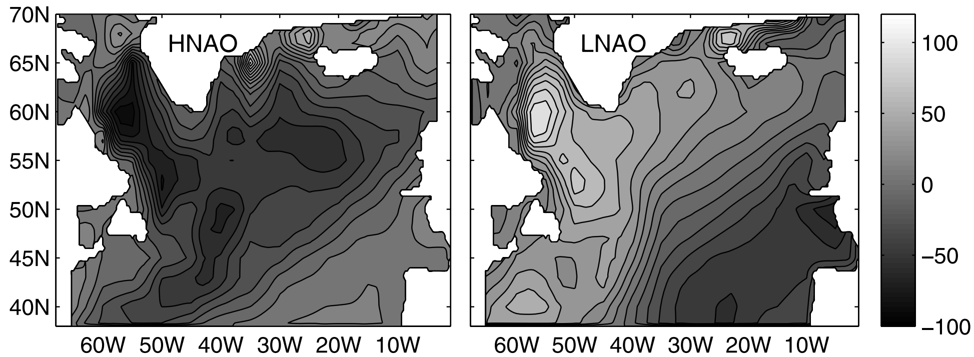

Figure 1. Heat flux anomalies (Wm^{-2}) averaged over December–February (NCEP/NCAR). HNAO anomalies are synthesized from three extreme high-NAO years (1989, 1990, and 1995) and LNAO anomalies from three extreme low-NAO years (1969, 1996, and 1963).

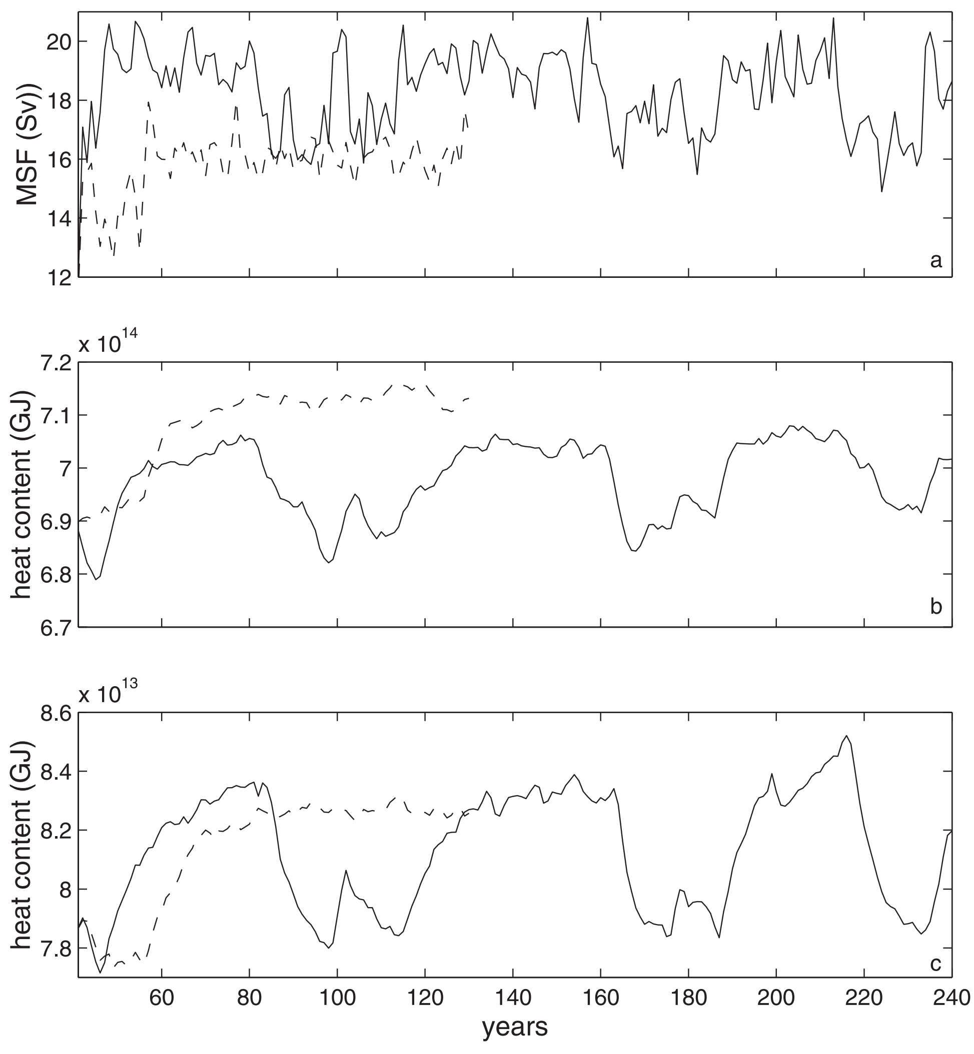

Figure 2. Time series of (a) meridional stream function at 55°N; (b) domain-averaged heat content; and (c) heat content in eastern basin (43–52°N 12–24°W) for HNAO (solid line) and LNAO (dashed line).

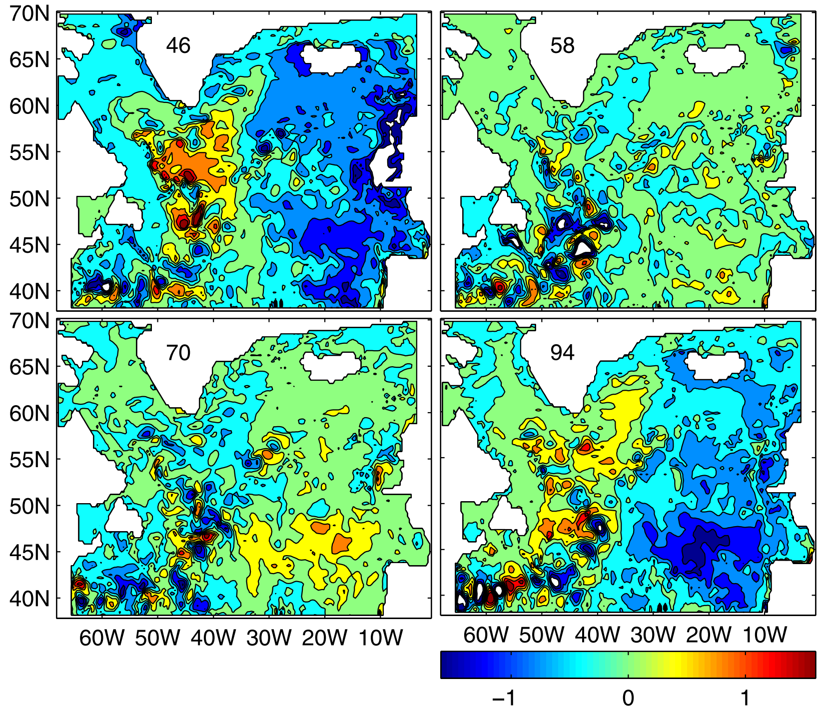

Figure 3. Snapshots of annual temperature anomalies (°C) in HNAO (with respect to annual temperature mean over the entire integration period: years 41–240) averaged over the upper 658 m (model levels 1–14) during a full oscillation (years 46–94), shown by four model years.

which feed back on the main currents (the NAC). In LNAO, heat transports are consistently directed northeastward after the spin-up, giving rise to a stable equilibrium situation.

[13] The average heat content along the southern boundary displays similar patterns to the basin average heat content, i.e., multidecadal variability in HNAO while approaching an equilibrium in LNAO, albeit with a slight lag. The observed variability in domain averaged heat content in HNAO is thus partly compensated for by variability in the inflow, but is driven by changes produced in the model interior. Note that the same open southern boundary conditions are implemented in both HNAO and LNAO.

3.2. Anomalous NAC Pathway

[14] We focus on the NAC's evolving patterns in the region 46–52°N 33–44°W by examining the model sea surface height (SSH) field. These variations are demonstrated using the SSH isolines of 0 and −5 cm. We plot the isolines for years 46, 70 and 94 for HNAO and for a mean of 51 years (years 80–130) for LNAO (Figure 4). In HNAO, a more westward NAC axis is evident, accompanied by low-frequency shifts of pathway between northwest (years 46 and 94) and northeast (year 70). Similar switches also appear in the subsequent oscillation. We do not observe a north-south NAC shift [*Bersch*, 2002], possibly because of the longitude of the modeled Northwest Corner and the proximity of this region to the model southern boundary. In LNAO, the NAC manifests a stable, more eastward (in contrast to HNAO and consistent with *Bersch* [2002]) and then northeast oriented pathway, without interdecadal changes.

[15] *Bersch* [2002] documented changes in the upper layer circulation along Greenland-Ireland and Newfoundland-France sections when the NAO index dropped significantly in 1996 and 1997. He suggested a contraction of the subpolar gyre with a westward shift accompanied by an eastward shift of the Subarctic Front (SAF) was observed in the Iceland and the Newfoundland Basins. This feature seems consistent with our LNAO experiment. A comparison of salinity in the upper 1000 m versus longitude in LNAO against an experiment under climatological forcing [*Yang*, 2005] (not shown) shows the occurrence of higher-salinity water in the Iceland Basin, Rockall Trough and western European Basin, as well as lower-salinity water in the Irminger Basin during extreme low-NAO years. A further comparison of salinity on σ_0 = 27.2 and σ_0 = 27.5 in LNAO indicates an increased northward spreading of warm, saline, and less dense Subpolar Mode Water and Mediterranean Water, as suggested by *Bersch* [2002, Figures 3 and 5].

[16] In conclusion, the multidecadal variability observed in the HNAO experiment is the consequence of the interaction of Labrador Sea deep convection and the change in the NAC pathway. That is, high-NAO forcing leads to cold conditions over Labrador Sea, leading to increased convection; the change in the density field then leads to a pathway shift in the NAC toward the northwest; this increases the oceanic heat flux into the Labrador Sea and hence warms the Labrador Sea, reducing the convection; the change in

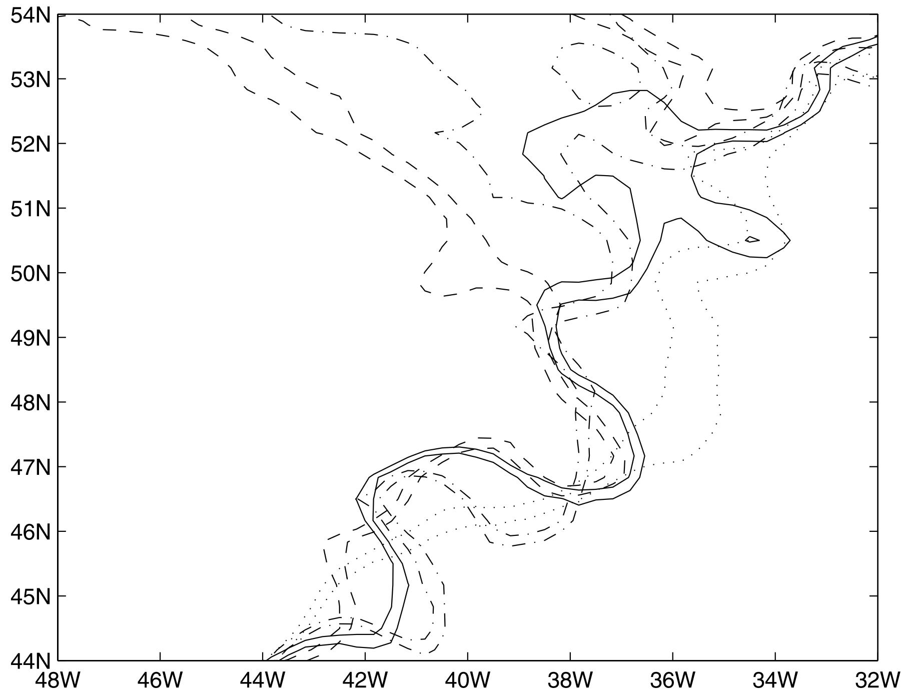

Figure 4. 0 and −5 cm sea surface height contours for years 46 (dashed line), 70 (solid line), and 94 (dash-dot line) in HNAO and for mean of 51 years (years 80–130) (dotted line) in LNAO, output from our model.

the density field shifts the pathway of the NAC back toward the northeast; the reduced heat flux into the Labrador Sea allows the Labrador Sea to cool and convection to resume. This closes the loop.

4. Discussion

[17] Using a hierarchy of increasingly complex model configurations under prescribed, idealized heat flux forcing, *te Raa et al.* [2004] proposed that interdecadal variability in the North Atlantic was caused by an internal oceanic mode, which can be identified by the following features; that is, northwestward propagation of large-scale temperature anomalies gives rise to a phase difference between meridional and zonal temperature gradient anomalies; anomalous east-west temperature difference leads anomalous north-south temperature difference; and anomalous east-west/north-south temperature differences lead to associated density anomalies that produce anomalous meridional/zonal overturning. We feel that such a mechanism can be essentially seen in our more "realistic" situations although such a process involves an interaction between LSW formation and the NAC path in our case. If we reexamine our temperature anomalies by only looking at Figure 3, our temperature anomalies also propagate northwestward in an alternating form of negative and positive phases, similar to te Raa et al.'s Figure 1. Figure 5 gives anomalous east-west/north-south temperature differences, which are calculated as defined by

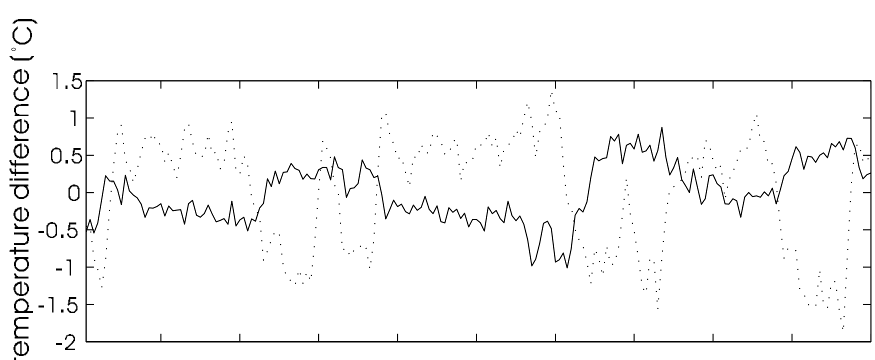

Figure 5. Time series of zonally averaged north-south temperature difference $\Delta\ T_{N-S}$ (solid line) and meridionally averaged east-west temperature difference $\Delta\ T_{E-W}$ (dotted line), averaged over the upper 1415 m.

te Raa et al. [2004] (their equations (3) and (4)). The anomalous east-west temperature difference is leading the anomalous north-south temperature difference at interdecadal scale. Note that we do not calculate anomalous overturning streamfunctions since the condition of horizontal nondivergence is not met for defining the zonal overturning streamfunction due to the nonzero velocity at the open southern boundary. Therefore we delineate this process: the intensive heat loss over Labrador Sea and subsequent vertical convection during the high-NAO phase (initiation of anomalous east-west temperature difference) increases the meridional heat flux via an anomalous NAC north-south transport, leading to anomalous north-south temperature difference, which then produces anomalous zonal heat flux via an anomalous NAC east-west transport; thus the internal low-frequency mode.

[18] *Curry and McCartney* [2001, Figures 4 and 8] plot the potential energy anomaly (PEA) distributions for years 1990–1997 (high-NAO phase) and 1965–1974 (low-NAO phase), as well as the difference between these two periods. A north-south dipole with increased (uplifted) subtropical PEA and decreased (deepened) subpolar PEA appears around Mann Eddy-Grand Banks and an east-west dipole with decrease to the west and increase to the east, north of Mann Eddy. These dipoles would seem to suggest strength variations of the NAC, and possibly also reflect an axis shift [*Curry and McCartney*, 2001], after a transit of the NAO index from low to high phase although the authors could not further verify it because of a lack of data. Our results support such a hypothesis.

5. Summary

[19] We examine the impact of forcing a regional eddy-permitting ocean model of the subpolar North Atlantic with anomalous buoyancy fluxes corresponding to extreme high- and low-NAO forcing for extended periods of time. It has been speculated that the MWP and the LIA are associated with several centuries during which the NAO is predominantly in one phase [*Bruckner and Mackensen*, 2006]. Consistent with a colder Atlantic sector during the LIA, we find weakened MOC in our low-NAO simulations and an enhanced MOC in our high-NAO simulations (potentially consistent with the MWP).

[20] Our results also show multidecadal variability with a period of 70 years in LSW formation and MOC in our high-NAO simulations, in comparison to an equilibrium like circulation under anomalous low NAO related forcing. A less stable state is therefore suggested to occur under the high-NAO phase and it appears to be brought about by oceanic internal process, i.e., changes in the NAC strength and pathway but excited by Labrador Sea deep convection, triggered by intensive heat loss during winter. The oscillation in both Labrador Sea convection and the NAC path then self-maintains without a change in surface forcing (such as a switch to low-NAO-related forcing).

[21] *Mann* [2003] argues against a significant MWP, and furthermore suggests the warming that occurred did not occur synchronously in all regions around the North Atlantic. *Crowley and Lowery* [2000] showed that western Greenland exhibited anomalous warmth around the start and end of the MWP, with cold conditions during the intervening period, which potentially overlapped with a period of warmth in Scandinavia. Such variability is consistent in principle with our variability during the high-NAO simulations, whereby the main pathway of the NAC (and thus its heat transport) shifted from into the Labrador Sea (and thus western Greenland) to the eastern Atlantic (and thus potentially Scandinavia). Although finding paleoclimatic evidence showing a clear signature of greater North Atlantic variability during a warm state (as an analogy to the MWP and/or a period of extended high NAO, as in our experiment) has proved difficult, we note that *Gil et al.* [2006] present two cores from Europe (Skaggerak, Tagus pro-delta) that seem to show enhanced variability during the MWP as compared to the LIA.

[22] We also wonder whether our modeled variability can be related to the AMO. Our period of 70 years is surprisingly close to the observed 65 year period [*Delworth and Mann*, 2000] as well as to simulated 70–120 periods in the Hadley Centre coupled climate model [*Knight et al.*, 2005]. They [*Knight et al.*, 2005] also find a link between their oscillation and the strength of the MOC and demonstrate that the deep ocean is needed to produce the AMO. If our oscillation during the high-NAO run is indeed related to the AMO, we confirm the link between the AMO and the MOC, as well as the importance of oceanic processes. Although quite speculative, we would furthermore suggest that even if the AMO is primarily an oceanic mode, it may be even lower-frequency climatic variability (the creation of a regime with extended NAO forcing) that excites this variability, without touching on the questions of what generates the extended periods with the NAO predominantly in one sign, or how/if the internal variability is damped/changed when the forcing changes. With the speculation that a warming world may be associated with a more positive NAO [*Fyfe et al.*, 1999], these results, if shown to be robust, suggest that further study is needed to understand if a warmer world will be a less stable world. As well these results emphasize the importance of understanding how the ocean circulation will evolve in such a situation.

[23] **Acknowledgments.** This work is part of Ph.D. thesis of Duo Yang and was funded by CFCAS (GR-019) and NSERC grants awarded to PGM. We would like to thank the anonymous reviewers for their useful thoughts that helped to improve this manuscript.

References

Appenzeller, C., T. F. Stocker, and M. Anklin (1998), North Atlantic Oscillation dynamics recorded in Greenland ice cores, *Science*, *282*, 446–449.

Bersch, M. (2002), North Atlantic Oscillation-induced changes of the upper layer circulation in the North Atlantic Ocean, *J. Geophys. Res.*, *107*(C10), 3156, doi:10.1029/2001JC000901.

Bruckner, S., and A. Mackensen (2006), Deep-water renewal in the Skagerrak during the last 1200 years triggered by the North Atlantic Oscillation: Evidence from benthic foraminiferal delta O-18, *Holocene*, *16*, 331–340.

Clarke, M. L., and H. M. Rendell (2006), Effects of storminess, sand supply and the North Atlantic Oscillation on sand invasion and coastal dune accretion in western Portugal, *Holocene*, *16*, 341–355.

Cook, E. R., R. D. D'Arrigo, and K. Briffa (1998), A reconstruction of the North Atlantic Oscillation using tree-ring chronologies from North America and Europe, *Holocene*, *8*, 9–17.

Cook, E. R., R. D. D'Arrigo, and M. E. Mann (2002), A well-verified, multiproxy reconstruction of the winter North Atlantic Oscillation index since A.D. 1400, *J. Clim.*, *15*, 1754–1764.

Crowley, T. J., and T. S. Lowery (2000), How warm was the medieval warm period?, *AMBIO: A J. Human Environ.*, *29*, 51–54.

Curry, R. G., and M. S. McCartney (2001), Ocean gyre circulation changes associated with the North Atlantic Oscillation, *J. Phys. Oceanogr*, *31*, 3374–3400.

Curry, R. G., M. S. McCartney, and T. M. Joyce (1998), Oceanic transport of subpolar climate signals to mid-depth subtropical waters, *Nature*, *391*, 575–577.

Delworth, T. L., and M. E. Mann (2000), Observed and simulated multidecadal variability in the Northern Hemisphere, *Clim. Dyn.*, *16*, 661–676.

Eden, C., and J. Willebrand (2001), Mechanism of interannual to decadal variability of the North Atlantic circulation, *J. Clim.*, *14*, 2266–2280.

Fyfe, J. C., G. J. Boer, and G. M. Flato (1999), The Arctic and Antarctic oscillations and their projected changes under global warming, *Geophys. Res. Lett.*, *26*, 1601–1604.

Gil, I. M., F. Abrantes, and D. Hebbeln (2006), The North Atlantic Oscillation forcing through the last 2000 years: Spatial variability as revealed by high-resolution marine diatom records from N and SW Europe, *Mar. Micropaleontol.*, *60*, 113–129.

Hurrell, J. W. (1995), Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation, *Science*, *269*, 676–679.

Hurrell, J. W., Y. Kushnir, G. Ottersen, and M. Visbeck (2003), An Overview of the North Atlantic Oscillation, in *The North Atlantic Oscillation: Climatic Significance and Environmental Impact*, *Geophys. Monogr. Ser.*, vol. 134, edited by J. W. Hurrell et al., pp. 1–35, AGU, Washington, D. C.

IPCC (2001), *IPCC Third Assessment Report: Climate Change 2001*, World Meteorol. Org., Geneva.

Jones, P. D., T. J. Osborn, and K. R. Briffa (2001), The evolution of climate over the last millennium, *Science*, *292*, 662–667.

Josey, S. A., E. C. Kent, and P. K. Taylor (1998), The Southampton Oceanography Centre (SOC) Ocean-Atmosphere Heat, Momentum and Freshwater Flux Atlas, *Rep. 6*, Nat. Oceanogr. Cent., Southampton, Southampton, UK.

Kalnay, E., et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bull. Am. Meteorol. Soc.*, *77*, 437–471.

Keigwin, L. D., and R. S. Pickart (1999), Slope water current over the Laurentian Fan on interannual to millennial time scales, *Science*, *286*, 520–523.

Knight, J. R., R. J. Allan, C. K. Folland, M. Vellinga, and M. E. Mann (2005), A signature of persistent natural thermohaline circulation cycles in observed climate, *Geophys. Res. Lett.*, *32*, L20708, doi:10.1029/2005GL024233.

Luterbacher, J., et al. (2002), Extending North Atlantic Oscillation reconstructions back to 1500, *Atmos. Sci. Lett.*, *2*, 114–124.

Mann, M. E. (2003), Medieval climatic optimum, in *The Earth System: Physical and Chemical Dimensions of Global Environmental Change, Encyclopedia of Global Environmental Change*, vol. 1, edited by M. C. MacCracken and J. S. Perry, pp. 514–516, John Wiley, Hoboken, N. J.

Myers, P. G. (2002), SPOM: A regional model of the sub-polar North Atlantic, *Atmos. Ocean*, *40*, 445–463.

Proctor, C. J., A. Baker, W. L. Barnes, and M. A. Gilmour (2000), A thousand year speleothem proxy record of North Atlantic climate from Scotland, *Clim. Dyn.*, *16*, 815–820.

Proctor, C. J., A. Baker, and W. L. Barnes (2002), A three thousand year record of North Atlantic climate, *Climate Dynamics*, *19*, 449–454.

te Raa, L. A., J. Gerrits, and H. A. Dijkstra (2004), Identification of the mechanism of interdecadal variability in the North Atlantic ocean, *J. Phys. Oceanogr.*, *34*, 2792–2807.

Treguier, A. M., S. Theetten, E. P. Chassignet, T. Penduff, R. Smith, L. Talley, J. Q. Beismann, and C. Boning (2005), The North Atlantic subpolar gyre in four high-resolution models, *J. Phys. Oceanogr.*, *35*, 757–774.

Trenberth, K. E., W. G. Large, and J. G. Olson (1990), The mean annual cycle in global ocean wind stress, *J. Phys. Oceanogr.*, *20*, 1742–1760.

Worthington, L. V. (1976), *On the North Atlantic circulation*, 110 pp., The Johns Hopkins Univ. Press, Baltimore, Md.

Yang, D. (2005), Impact of anomalous surface forcing on the sub-polar North Atlantic: Water mass formation and circulation, Ph.D. thesis, 131 pp., Univ. of Alberta, Edmonton, Alberta, Canada.

P. G. Myers, Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alberta, Canada, T6G 2E3. (pmyers@ualberta.ca)

D. Yang, Department of Oceanography, Dalhousie University, Halifax, Nova Scotia, Canada, B3H 4J1. (duo.yang@phys.ocean.dal.ca)