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A review of North Atlantic modes of natural variability and their driving mechanisms

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[1] This paper reviews three modes of natural variability that have been identified in the North Atlantic Ocean, namely, the North Atlantic Oscillation (NAO), the Atlantic Multidecadal Oscillation (AMO) and the Atlantic Meridional Mode (AMM). This manuscript focuses on the multidecadal fluctuations of these three modes. A range of different mechanisms to initiate phase reversals in these modes on multidecadal timescales has been suggested previously. We propose a systematic grouping of these mechanisms into three types that involve, respectively, (1) the dependency of the Atlantic thermohaline circulation (THC) on salinity, (2) the sensitivity of the THC to changes in ocean heat transport and (3) the dependency of the NAO to changes in the Atlantic meridional temperature gradient. Some new density data is also provided, demonstrating physical links between the THC and the AMO.

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1. Introduction

[2] There is a growing body of literature on decadal and multidecadal modes of natural variability in the Atlantic and on the climatic impacts associated with these modes, including variations in North Atlantic tropical cyclone activity [*Goldenberg et al.*, 2001; *Kossin and Vimont*, 2007; *Klotzbach and Gray*, 2008]. We here understand a mode of variability to be a large-scale climate pattern characterized by a particular set of atmospheric/oceanic conditions. Three interrelated modes with variability in the decadal and multidecadal spectrum in addition to shorter-term variations have been identified [e.g., *Marshall et al.*, 2001a]: the North Atlantic Oscillation (NAO) [*Van Loon and Rogers*, 1978; *Lamb and Peppler*, 1987], the Atlantic Multidecadal Oscillation (AMO) [*Bjerknes*, 1964; *Delworth et al.*, 1993; *Gray et al.*, 1997], and the Atlantic Meridional Mode (AMM) [*Servain*, 1991; *Xie*, 1999; *Vimont and Kossin*, 2007]. We will focus on the variability of these three modes on the multidecadal timescale.

[3] A range of different mechanisms that potentially underlie these modes and initiate phase reversals has been suggested previously. This paper augments previous research through a systematic grouping of these mechanisms into three types that involve, respectively, 1) the dependency of the Atlantic thermohaline circulation (THC) on salinity, 2) the sensitivity of the THC to changes in ocean heat transport and 3) the dependency of the NAO on changes in the Atlantic meridional temperature gradient.

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[4] Section 2 gives a brief overview of the AMO, AMM and NAO. In section 3, we begin by reviewing the evolution of the theoretical framework in earlier work on Atlantic atmosphere-ocean variability, highlighting that the main mechanisms associated with the Atlantic modes have been known for a number of decades. Employing the above grouping into three categories, we then provide a systematic overview of mechanisms from which the three modes are proposed to arise and interact and through which they influence other phenomena of interest. Section 4 discusses multidecadal variability in the observed data. We also present some new density data and discuss documented evidence of changes in a variety of parameters in the context of the proposed phases of the AMO and AMM since 1878. Section 5 summarizes the manuscript and provides ideas for future work.

2. The Three Atlantic Modes and Their Interrelationships

2.1. North Atlantic Oscillation

[5] The North Atlantic Oscillation (NAO) is the synchronous variation of the pressure gradient between the Icelandic Low and the Azores High on timescales from daily to multidecadal [*Van Loon and Rogers*, 1978]. A positive NAO index signifies an increased pressure gradient and stronger westerlies across the Atlantic. The positive (negative) phase of the NAO is also associated with a tripole pattern of cold (warm) sea surface temperature (SST) anomalies in the subpolar North Atlantic, a warm (cold) anomaly in the West Atlantic between 20 and 45°N that narrows toward the east, and a cold (warm) anomaly between 0 and 30°N in the East Atlantic [*Van Loon and Rogers*, 1978; *Marshall et al.*, 2001a].

[6] More recently, *Thompson and Wallace* [1998] have documented a more coherent, large-scale mode of variability termed the Arctic Oscillation (AO) that involves shifts in

mass between the polar and subpolar Northern Hemisphere latitudes. The NAO can be viewed as the North Atlantic regional manifestation of the AO [*Marshall et al.*, 2001a]. The NAO and AO are closely related, with correlations between the two indices exceeding 0.70 on monthly timescales during the winter months [*Thompson and Wallace*, 1998]. Given our focus on the Atlantic sector, we primarily discuss fluctuations in the NAO.

2.2. Atlantic Multidecadal Oscillation

[7] The Atlantic Multidecadal Oscillation (AMO) is associated with basin-wide SST and SLP fluctuations. For the positive AMO phase, this is sometimes presented as an almost uniform warming of the North Atlantic [*Enfield et al.*, 2001]. The original work on the AMO which we build on here associates the positive AMO phase with a pattern of horseshoe-shaped SST anomalies in the North Atlantic with pronounced warming in the tropical and parts of the eastern subtropical North Atlantic, an anomalously cool area off the U.S. East Coast, and warm anomalies surrounding the southern tip of Greenland [*Bjerknes*, 1964; *Kushnir*, 1994]. This definition of the AMO has been continued in more recent work [*Gray et al.*, 1997; *Goldenberg et al.*, 2001; *Klotzbach and Gray*, 2008]. This SST pattern is suggested to be linked to variations in the strength of the Atlantic THC, the density-driven component of the overturning circulation that transports approximately 10^{15}W of heat into the higher latitudes of the North Atlantic [*Toggweiler and Key*, 2001]. The THC is driven by the sinking of salty water that cools and becomes denser, forming North Atlantic Deep Water (NADW) in the higher latitudes of the North Atlantic [*Dickson and Brown*, 1994; *Marshall and Schott*, 1999].

[8] Due to the strong salinity/density relationship at water temperatures near freezing, the density of NADW can change significantly with small variations in salinity content. Downwelling is also affected by changes in the intensity of westerly winds [*Bjerknes*, 1964]. Long-term salinity variations, as documented over this past century [*Smed*, 1943; *Dooley et al.*, 1984; *Dickson et al.*, 1988; *Belkin et al.*, 1998] are expected to affect the strength of the THC, with its potential SST impacts resulting in the multidecadal pattern associated with the AMO [*Bjerknes*, 1964; *Delworth et al.*, 1993; *Gray et al.*, 1997; *Goldenberg et al.*, 2001].

2.3. Atlantic Meridional Mode

[9] The Atlantic Meridional Mode (AMM) is also known historically as the Atlantic Dipole or Interhemispheric Mode [*Servain*, 1991; *Xie and Philander*, 1994; *Carton et al.*, 1996] or the tropical Atlantic gradient mode [*Chiang et al.*, 2002]. It is characterized by variations in SST and SLP between the tropical Atlantic north and south of the Intertropical Convergence Zone (ITCZ) [*Vimont and Kossin*, 2007]. A positive AMM is associated with an anomalously strong north-south tropical Atlantic SST gradient and an anomalously weak SLP gradient. Variations in these gradients do not necessarily imply that SST and SLP north and south of the ITCZ covary [*Xie and Carton*, 2004]. The gradients during the positive AMM phase induce an anomalous cross-equatorial southerly flow that diverts into southeasterly (southwesterly) winds south (north) of the equator. These anomalous winds weaken the northern tropical Atlantic trades while strengthening the southern tropical Atlantic trades. The ITCZ, which is sensitive to variations in the cross-equatorial SST gradient [*Chiang et al.*, 2002], follows the northward shift of SST patterns and trade wind convergence.

2.4. Interrelationship Between the Three Modes

[10] The cross-equatorial pattern associated with the AMM and the SST, SLP and wind patterns associated with the AMO have been viewed as one overall phenomenon that stretches from the high latitudes to the tropics [*Gray*, 1990; *Gray et al.*, 1997; *Xie and Tanimoto*, 1998] or as two distinct but closely related modes [*Kossin and Vimont*, 2007]. Generally, the AMO is thought to excite the AMM on longer timescales [*Vimont and Kossin*, 2007]. On shorter timescales, positive AMM conditions may also arise in association with El Niño events and/or negative NAO events [*Czaja et al.*, 2002]. These shorter time variations may be captured best when the AMO and AMM are treated as two distinct modes [*Kossin and Vimont*, 2007].

[11] The AMO and AMM are also closely related to the NAO on (multi)decadal time scales [*Xie and Tanimoto*, 1998; *Marshall et al.*, 2001a]. Long-term positive (negative) phases of the NAO coincide with the negative (positive) phase of the AMO and AMM, generally with a lag of several years [e.g., *Rogers*, 1985; *Robertson et al.*, 2000; *Zhang and Vallis*, 2006]. This relationship appears to be part of the negative (balancing) feedbacks discussed in detail in section 3 through which the AMO and NAO switch between long-term predominantly positive and negative phases. In particular, the NAO depends on the North Atlantic meridional temperature and pressure gradient, which in turn lessens (increases) as the North Atlantic warms (cools) with the positive (negative) AMO. The NAO in turn is suggested to affect the THC through its effects on sea ice export and heat loss in the areas of NADW formation.

3. Mechanisms Underlying Atlantic Multidecadal Variability

[12] The AMO has recently been proposed to be a construct of aerosol forcing in the North Atlantic and large-scale warming primarily driven by greenhouse gas increases [*Mann and Emanuel*, 2006]. However, the ocean-atmosphere variations and climatic effects associated with the AMO are quite diverse, including for instance changes in ocean currents and subsurface temperatures [*Bjerknes*, 1964; *Zhang*, 2007, 2008; *Zhang and Vallis*, 2006] as well as impacts outside the Atlantic [*Folland et al.*, 1986; *Gray et al.*, 1997; *Enfield et al.*, 2001; *Sutton and Hodson*, 2005; *Zhang et al.*, 2007; *Dima and Lohmann*, 2007]. Aerosol forcing, as assumed by *Mann and Emanuel* [2006], would not affect the subsurface as rapidly and distinctly as the ocean surface [*Zhang*, 2007].

[13] More importantly, while SSTs in the tropical North Atlantic from 0 to 30°N and in the far north between 40 and 70°N can be used as an indicator of the state of the AMO [*Goldenberg et al.*, 2001], the signal of the AMO was originally discovered as near-global in extent [*Folland et al.*, 1986]. Later studies confirmed that SST anomalies extend outside the North Atlantic [*Gray et al.*, 1997; *Enfield*

and Mestaz-Nuñez, 1999]. Decomposing tropical North Atlantic SST into a global trend and a residual component that is assumed to contain the AMO signal, as proposed by *Mann and Emanuel* [2006], is therefore problematic, as part of the AMO signal will be contained in the global component.

[14] The appearance of multidecadal periodicity in long-term Atlantic modes due to internal mechanisms necessitates the existence of three conditions: positive feedback mechanism(s) that act to amplify variations in Atlantic SST and other parameters and allow each state of the mode to persist, negative feedback mechanism(s) that are able to act against operating positive feedbacks toward a shift to the respective other state of the mode, and mechanism(s) of delay or memory [*Suarez and Schopf*, 1988].

[15] The main variables that change with the phase changes of the AMO, AMM and NAO are SST or oceanic heat content, wind patterns and salinity. Accordingly, mechanisms discussed in the literature can be grouped into three types of variations: (1) changes in air-sea heat flux and SST, (2) changes in momentum flux and wind patterns, and feedbacks between wind and SST patterns, and (3) changes in water flux and buoyancy flux and their effects on the North Atlantic salinity budget. The first two types of variations tend to involve atmospheric forcing of the ocean, while in the third case the ocean may be forcing atmospheric changes through long-term variations in the THC [*Grötzner et al.*, 1998]. Before discussing the role of these variations for the Atlantic modes in more detail, we will give a brief overview of the evolution of the theoretical framework.

3.1. Evolution of the Theoretical Framework

[16] Observations of coordinated long-term variations in North Atlantic SST, SLP, salinity and wind patterns related to the three Atlantic modes have been documented for almost a century [*Helland-Hansen and Nansen*, 1917; *Defant*, 1924; *Walker*, 1924; *Wüst*, 1935; *Smed*, 1943]. *Helland-Hansen and Nansen* [1917] proposed that variations in the strength of the midlatitude westerlies and tropical trade winds that they documented were induced by concurrent SST changes. *Walker* [1924], *Defant* [1924], and *Wiese* [1924] discussed the principal mechanisms of the NAO and their effect on midlatitude wind patterns, winter temperatures in Europe and Greenland, Arctic sea ice, and preferred cyclone tracks. *Defant* [1924] proposed the existence of two alternating states of the North Atlantic atmosphere-ocean environment, corresponding respectively to the positive and negative phases of the NAO.

[17] In the first four decades of the 20th century, studies began to discuss variations of category 3 above, namely the relationship between NADW formation and changes in salinity, temperature and buoyancy [e.g., *Helland-Hansen and Nansen*, 1917; *Wüst*, 1935; *Smed*, 1943]. *Smed* [1943] documented an increase in salinity from the period 1902–1917 to the period 1919–1939 that coincided with a warming of SSTs. This was followed by *Bjerknes'* [1961, 1964] seminal analysis that combined heat, momentum and buoyancy flux mechanisms into a framework for Atlantic air-sea interaction. Bjerknes presented detailed SST, SLP and wind data for the period 1880–1938 (with some additional data extending to 1960) in support of his hypothesis of short-term and long-term variability. In his framework, short-term variability arises due to the effect of the average strength of the westerlies and the trades on the net heat loss of the ocean to the atmosphere. Long-term variability arises due to systematic variations in the strength of the cross-equatorial and poleward heat transport via the THC. Enhanced (reduced) evaporation due to stronger (weaker) westerlies will lead to enhanced (reduced) upwelling and subsequent cooling (warming) and increased (decreased) density.

[18] In the 1950s and 1960s, studies explored mechanisms of type 2 in the cross-equatorial tropical environment, focusing on the relationship of variations in the cross-equatorial SST gradient with the meridional excursion of the monsoon belt, the strength of the trades [*Kraus*, 1956; *Namias*, 1963] and subtropical rainfall [*Kraus*, 1958]. *Kraus* [1958] noted long-term shifts in the Northern Hemisphere climate around 1900 and 1930, marked respectively by a reduction (increase) in the northward migration of the monsoon belt and subtropical summer rainfall, and an intensification (weakening) of the trades.

[19] Subsequent work expanded these arguments [e.g., *Lamb*, 1966; *Hastenrath*, 1976; *Hastenrath and Heller*, 1977; *Kraus*, 1977; *Lamb*, 1978; *Nicholson*, 1981], establishing for instance the development of a cross-equatorial flow into the Northern Hemisphere during periods of anomalously warm North Atlantic SSTs [*Hastenrath and Lamb*, 1977] that will further warm the eastern portion of the subtropical Atlantic through enhanced downwelling. Studies at that time also began to report correlations between Atlantic TCs and some of the identified variations [*Namias*, 1954; *Hastenrath*, 1976]. In particular, positive correlations with low subtropical SLP [*Brennan*, 1935] and with enhanced Sahel rainfall and a more northerly positioned ITCZ were reported [*Hastenrath*, 1976]. By the 1980s and early 1990s, many central mechanisms underlying long-term Atlantic variability and their impacts on TCs and other climate parameters were known [*Moura and Shukla*, 1981; *Shapiro*, 1982; *Reed*, 1988; *Gray*, 1990; *Hastenrath*, 1990; *Servain*, 1991]. Since the early 2000s, model simulation studies have begun to reproduce the climatic impacts of the AMO [e.g., *Giannini et al.*, 2003; *Knight et al.*, 2006; *Zhang and Delworth*, 2006; *Zhang et al.*, 2007; *Knutson et al.*, 2007].

3.2. Current Understanding of Mechanisms for Long-Term Atlantic Variability

3.2.1. Positive Feedbacks Enabling Long-Term Atlantic Variability

[20] As observed by *Bjerknes* [1964], the intensity of the North Atlantic westerlies and the trade winds acts to reinforce existing SST anomalies in the midlatitude and tropical North Atlantic, respectively. This will further act to reinforce the existing phase of the AMO (in the case of the westerlies) [*Häkkinen*, 2000] and the AMM (in the case of the trades) [*Kossin and Vimont*, 2007]. The reinforcing feedback between the trades and SST has been formalized as the Wind Evaporation SST (WES) feedback [*Xie and Philander*, 1994]. This feedback arises as the cross-equatorial flow into the Northern (Southern) Hemisphere during the positive (negative) AMM phase weakens (strengthens) the North Atlantic trades. This will decrease (increase) evaporation over the tropical North Atlantic, thus reinforcing the weaker (stronger) trades. Due to the relatively narrow width

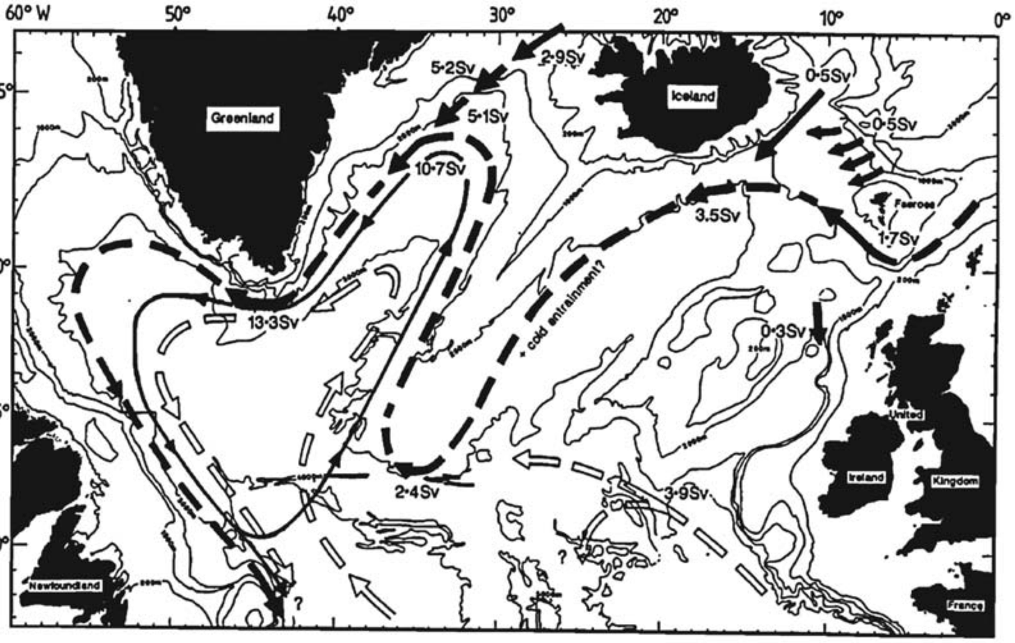

Figure 1. Quantification of deep water transport in the northern North Atlantic. Deep water is defined as denser than σ = 27.8. Figure taken from *Dickson and Brown* [1994].

of the North Atlantic, smaller wave numbers and a meridional mode rather than a zonal mode are favored [*Xie et al.*, 1999], making the AMM with its associated meridional WES feedback the leading mode of tropical ocean-atmosphere interactions [*Kossin and Vimont*, 2007].

[21] Positive feedbacks are also suggested to operate between convection and surface temperature-induced density changes in the Labrador Sea [*Marshall and Schott*, 1999; *Kuhlbrodt et al.*, 2001]. With enhanced (reduced) convection, surface heat fluxes increase (decrease), which enhances (weakens) convection further.

[22] A positive feedback may also exist between the NAO and heat transport in the subtropical gyre (STG) and North Atlantic Current (NAC). As reproduced in a modeling study by *Krahmann et al.* [2001] and an observational study by *Sutton and Allen* [1997], the observed warm SST anomaly along the U.S. East Coast during the negative AMO and positive NAO phase [*Bjerknes*, 1961, 1964] is gradually advected along the northern lobe of the Gulf Stream (GS). However, further research is needed to clarify the extent to which these NAO- and AMO-induced changes in heat transport in the GS are able to feedback on the atmosphere [*Frankignoul et al.*, 2001; *Krahmann et al.*, 2001].

3.2.2. Negative Feedbacks Enabling Long-Term Atlantic Variability

3.2.2.1. Variability of Convection due to Salinity and Temperature Changes

[23] Deep convection in the North Atlantic takes place in the Greenland-Iceland-Norwegian (GIN) Seas, the Labrador Sea, and to a lesser extent in the Irminger Sea and Mediterranean Sea. A quantification of the deep water flow through these areas is given by *Dickson and Brown* [1994] (Figure 1). Approximately 5.6 Sv of dense water enter the subpolar gyre (SPG) from the GIN Seas, composed of approximately 2.9 Sv from the Northwestern Stream between Iceland and Greenland and 2.7 Sv from the Northeastern Stream that passes through the Faroe Bank Channel east of Iceland. 3.2 Sv deep water are estimated to form in the Labrador Sea Water and join the Northeastern Stream via the SPG (Figure 1). Given further increases from entrainment and Antarctic Bottom Water, approximately 13.3 Sv flow southward from Cape Farewell [*Schmitz and McCartney*, 1993; *Dickson and Brown*, 1994].

[24] The depth and intensity of convection in the GIN and Labrador Seas depends sensitively on structural properties of the respective cyclonic gyres and the density of surface water relative to the underlying intermediate water. This density in turn changes with variations in salinity and temperature. Salinity variations can be caused by: 1) alterations in the strength of the THC and 2) alterations in freshwater influx into the core of the convective regions [*Aagaard and Carmack*, 1989; *Delworth et al.*, 1993; *Dickson et al.*, 1996; *Gray*, 1997].

[25] Generally, the high North Atlantic salinity is maintained through a net excess of evaporation over precipitation [*Gray et al.*, 1997]. When the THC runs more strongly, the time available for salinity buildup is shortened, leading to an eventual depletion of salinity and a switch to the negative AMO phase. This possibility has been explored in several theoretical and modeling studies [*Weaver et al.*, 1993; *Gray*, 1997; *Griffies and Bryan*, 1997; *Timmermann et al.*, 1998] (Figure 2). *Timmermann et al.* [1998] find that evaporation in the region off of Newfoundland and east of Greenland outweighs precipitation during the positive NAO phase. Convection strengthens after the positive NAO phase has peaked, followed by a strengthening of the THC with a short delay (Figure 2). In a simulation with the HadCM3 climate model, *Vellinga and Wu* [2004] find that tropical

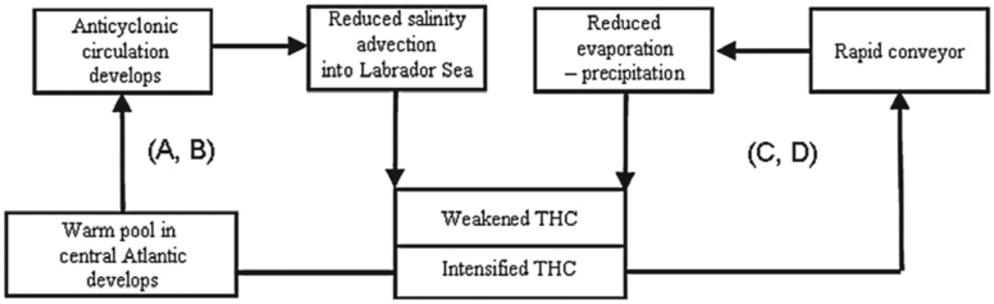

Figure 2. Variability of deep convection due to salinity and temperature changes as investigated in modeling and theoretical studies. References are: A, *Delworth et al.* [1993]; B, *Jungclaus and Haak* [2005]; C, *Gray* [1997]; and D, *Weaver et al.* [1993].

Atlantic precipitation increases due to the northward shift of the ITCZ in the strong phase of the THC. On long time scales, this freshened water is advected to higher latitudes, resulting in oscillations on a timescale of 70–200 years. A similar mechanism may underlie the 70–120 year oscillations simulated by *Knight et al.* [2005], who use the same model in a 1400 year run.

[26] Further approaches centered on multidecadal changes in salinity advection are explored by *Delworth et al.* [1993] with the fully coupled ocean-atmosphere model of the Geophysical Fluids Dynamics Laboratory (GFDL) and by *Jungclaus and Haak* [2005] in a fully coupled, non-flux-adjusted model. In these models, the THC fluctuates on approximately 50 year and 70–80 year time scales, respectively. The maximum density anomalies associated with a strong THC are located in the western Atlantic and in the Labrador Sea. When the THC is strong, an anomalous warm pool develops in the north central Atlantic due to enhanced warm advection. The associated anticyclonic circulation acts to reduce the amount of salinity being advected into the Labrador Sea [*Delworth et al.*, 1993; *Jungclaus and Haak*, 2005]. Differences between these two studies lie in the role played by Arctic-Atlantic interactions and the gyre circulation. In the model of *Delworth et al.* [1993], the changes in salinity transport via the anomalous circulation are due to the generally lower climatological salinity content in the eastern and central Atlantic relative to the western Atlantic. In the model of *Jungclaus and Haak* [2005], the anticyclonic circulation acts to restrict the freshwater export through the Fram Strait while allowing a buildup of salinity in the Arctic. In both models, a weakening of deep water formation in the Labrador Sea area and a phase switch of the THC follow (Figure 2). Thereafter, the north central Atlantic cools, leading to the development of a cyclonic circulation.

3.2.2.2. Variations in Labrador Sea Convection

[27] In recent decades, deep water formation in the Labrador Sea has been found to be closely linked to long-term trends in the NAO [*Dickson et al.*, 1988, 1996; *Bentsen et al.*, 2004], due to effects of the NAO on both temperature and salinity in the SPG region. The negative extreme of the NAO during the 1960s acted to first, minimize winter heat loss as storm tracks with the associated chilling winds shifted southward [*Dickson et al.*, 1996; *Curry and McCartney*, 2001] and second, to enable the export of large quantities of Arctic sea ice and fresh polar water (Figure 3). This is due to the development of anomalous northerly winds over the GIN Seas following the establishment of a persistent high pressure anomaly over Greenland during the negative NAO phase [*Dickson et al.*, 1988; *Aagaard and Carmack*, 1989; *Walsh and Chapman*, 1990; *Wohlleben and Weaver*, 1995; *Belkin et al.*, 1998] (Figure 3). Consequently, salinities near Iceland freshened to below the critical value of 34.7 psu that is necessary for deep water formation. Below this value, sea ice will form, enabling the preservation of the fresh, cold characteristics of the upper water. The great salinity anomaly (GSA) observed in the Labrador Sea in the late 1960s and circulating in the SPG thereafter has been attributed to this freshwater influx [*Dickson et al.*, 1988, 1996; *Belkin et al.*, 1998]. The decreased wind stress curl over the Labrador Sea further acted to reduce Ekman divergence and upwelling. This permitted the fresh surface water to spread into the convective center of the Labrador Sea [*Dickson et al.*, 1996] (Figure 3).

[28] *Dima and Lohmann* [2007] discuss Arctic sea ice export into the GIN and Labrador Seas as part of a hemispheric mechanism that potentially causes the observed patterns of the AMO (Figure 3). During the positive AMO phase, positive SSTs dominate the North Atlantic. This leads to the development of positive SLP anomalies over Greenland and the North Pacific, and negative anomalies over the North Atlantic and Siberia. The wave number 1 structure resulting from these anomalies forces northerly winds that in turn enhance sea ice export through Fram Strait. The resulting freshwater anomaly eventually leads to a shift to the negative AMO phase. Adjustment times at each step of this process result in an overall time frame of 30–35 years between sign changes [*Dima and Lohmann*, 2007].

[29] *Zhang and Vallis* [2006] explore the possible effects of GSA events on multidecadal variability in the North Atlantic with the GFDL global ocean general circulation model coupled to an atmospheric model. They show that the dipolar SST anomalies associated with the AMO—cooling south of Greenland and warming off the U.S. East Coast—can be triggered by the circulation of GSAs in the SPG, while greenhouse gas forcing combined with wind patterns are not able to reproduce the decadal variations. In their subsequent analysis of observational data back to 1900, they find a robust lag relationship between GSA events, the above SST dipole pattern and a shift of the NAO to a long-term positive phase. A similar oceanic forcing of the NAO

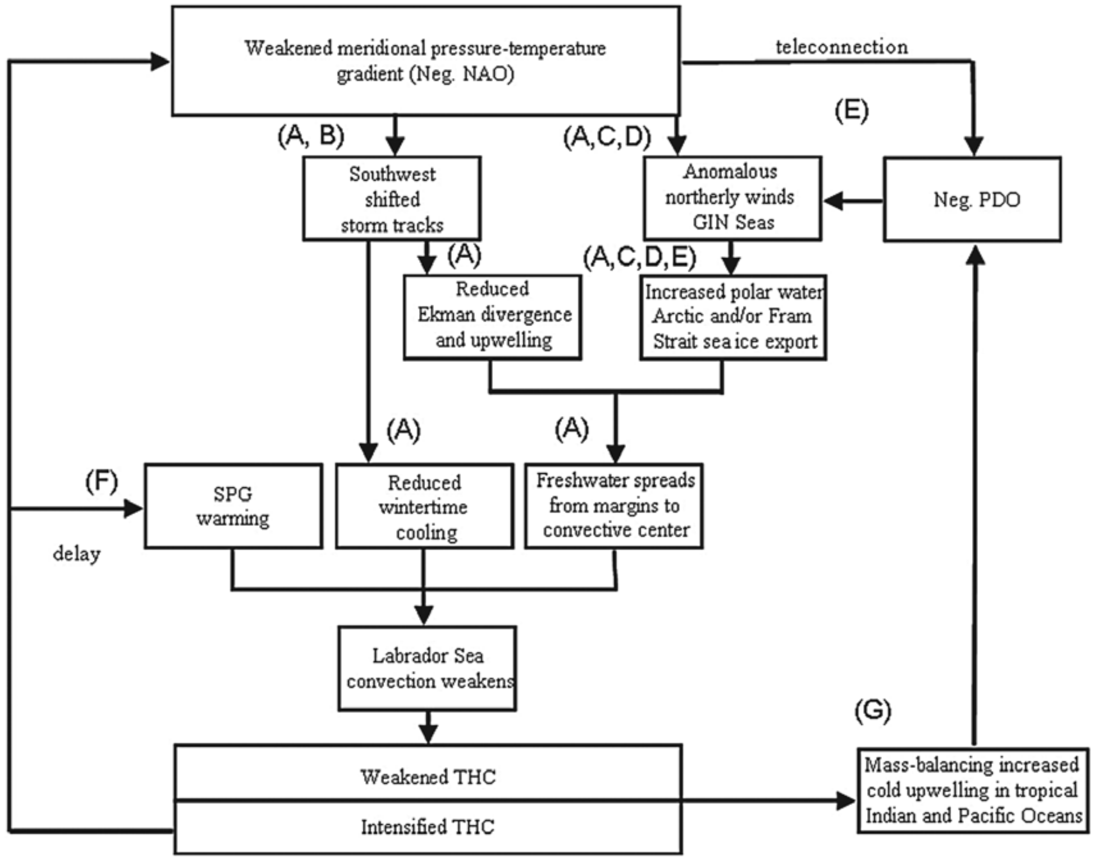

Figure 3. Variations in Labrador Sea convection and effect on the THC. References are: A, *Dickson et al.* [1988, 1996]; B, *Wohlleben and Weaver* [1995]; C, *Belkin et al.* [1998]; D, *Aagaard and Carmack* [1989]; E, *Dima and Lohmann* [2007]; F, *Häkkinen* [2000]; and G, *Gray* [1997].

has also been reproduced in modeling studies [*Rodwell et al.*, 1999; *Mehta et al.*, 2000], while other studies find only a weak response of the NAO to extratropical SST forcing [*Kushnir et al.*, 2002].

[30] It is expected that the GSA of the late 1960s was reinforced by local freshening as it circulated eastward along the SPG during the 1970s [*Dickson et al.*, 1988, 1996; *Belkin et al.*, 1998]. The first GSA may thus have reemerged as the observed 1980s anomaly [*Belkin et al.*, 1998]; alternatively the latter may have developed locally in the Labrador Sea as a consequence of severe sea ice conditions during the early 1980s [*Mysak and Manak*, 1989; *Belkin et al.*, 1998]. Detrimental effects of the 1980s anomaly on convection in the Labrador Sea were minimized, however, by the strengthening of the NAO from the early 1970s and the return of strong chilling wintertime wind stress curl over the Labrador Sea [*Dickson et al.*, 1996]. In the winter of 1971–72 and again in 1985, surface waters cooled sufficiently to cause the freshwater surface lid to overturn and deep convection to resume [*Curry and McCartney*, 2001].

[31] *Dickson et al.* [1988] estimate a salt deficit during the late 1960s equivalent to a fresh water excess of 2000 km^3. This is on a similar order as the annual sea ice export [*Aagaard and Carmack*, 1989] and the annual precipitation-evaporation in the GIN Seas [*Häkkinen*, 1999]. The effects of a salinity anomaly of this extent on the strength of the THC and the GS have been investigated in a number of modeling studies. The simulated reductions in the strength of the GS range from negligible effects to reductions between 30 and 60% (Table 1). Models with a more realistic initial GS strength generally find much greater impacts. *Gray* [1997] gives a theoretical estimate of a 25–35% reduction in the strength of the THC. *Holland et al.* [2001] estimate a 10% reduction in the THC on the basis of a global coupled ice-ocean-atmosphere model. *Schlosser et al.* [1991] find a 80% reduction in NADW formation during and after the time of the GSA based on tracer data and a diagnostic model of NADW formation.

[32] Based on hydrographic observations and on observational and proxy evidence of sea ice, storm and wind patterns, *Dickson et al.* [1996] hypothesize that these patterns of sea ice advection anomalies may be part of a regular pan-Atlantic pattern, with another possible convective maximum in the Labrador Sea during the early years of the positive AMO phase in the mid 1920s. In this framework, once a negative NAO phase has become fully established, Labrador Sea convection should weaken due to the combined effect of the NAO on salinity and winter heat loss. A recovery of convection should follow a subsequent strengthening of the NAO. It appears that the effect of enhanced sea ice advection involves a considerable delay, while the effect of a resumption in wintertime cooling may be more prompt [*Dickson et al.*, 1996; *Curry and McCartney*,

Table 1. Estimates of the Effect of Salinity Anomalies Comparable to the 1970s GSA on the Strength of the GS[a]

Study	Type of Model	Original GS (Sv)	GS Reduction (%)
Myers et al. [2005]	Diagnostic model: extension of *Greatbatch et al.* [1991]	118.75	28
Saenko et al. [2003]	Coupled global intermediate complexity model with sea ice	14.8	5
Häkkinen [1999]	Coupled ice-ocean model	40	20–30
Ezer et al. [1995]	Diagnostic model	30–40	60–75
Power et al. [1994]	Global ocean general circulation model	22	Insignificant
Greatbatch et al. [1991]	Diagnostic model	40–60	50

[a]The strength of the original GS given by *Myers et al.* [2005] is an average over the strongest pentads, and the strength of the original GS for *Häkkinen* [1999], *Ezer et al.* [1995], and *Greatbatch et al.* [1991] are scaled from their diagrams.

2001]. This suggests that positive AMO phases may have a duration exceeding that of negative phases, similar to what has been observed since the late 1800s [e.g., *Goldenberg et al.*, 2001].

3.2.2.3. Variations in GIN Sea Convection

[33] Observational [*Dickson et al.*, 1996] and modeling [*Bentsen et al.*, 2004] studies indicate that convection in the GIN and Labrador Seas in recent decades have been anticorrelated concurrently or correlated with a lag of more than a decade. At the beginning of the first salinity anomaly, convection in the GIN Seas was at a maximum [*Dickson et al.*, 1996]. During the passage of the first GSA, GIN Sea convection remained mostly shielded from freshwater influx by a strong polar front (Figure 4). In contrast to the situation in the Labrador Sea, convection in the GIN Seas is supported by the strong wintertime cooling and high wind stress curl prevailing during negative NAO conditions. Wind stress curl acts to strengthen the cyclonic gyre in the convective region and the frontal zones that surround it (Figure 4). Wind stress curl also acts to enhance surface layer divergence and the doming of isopycnals that bring high density water to the surface, promoting weak stratification that is required for dense surface water to be able to sink [*Marshall and Schott*, 1999] (Figure 4). Once the NAO had begun to strengthen, convection progressively weakened during the passage of the 1980s salinity anomaly [*Schlosser et al.*, 1991; *Dickson et al.*, 1996]. At that point, Labrador Sea convection had begun to resume, due to the effect of several years of a strengthening NAO.

[34] It is unclear whether, as suggested by some modeling and observational studies [*Delworth et al.*, 1993; *Deser and Blackmon*, 1993; *Häkkinen*, 1999; *Bailey et al.*, 2005] the

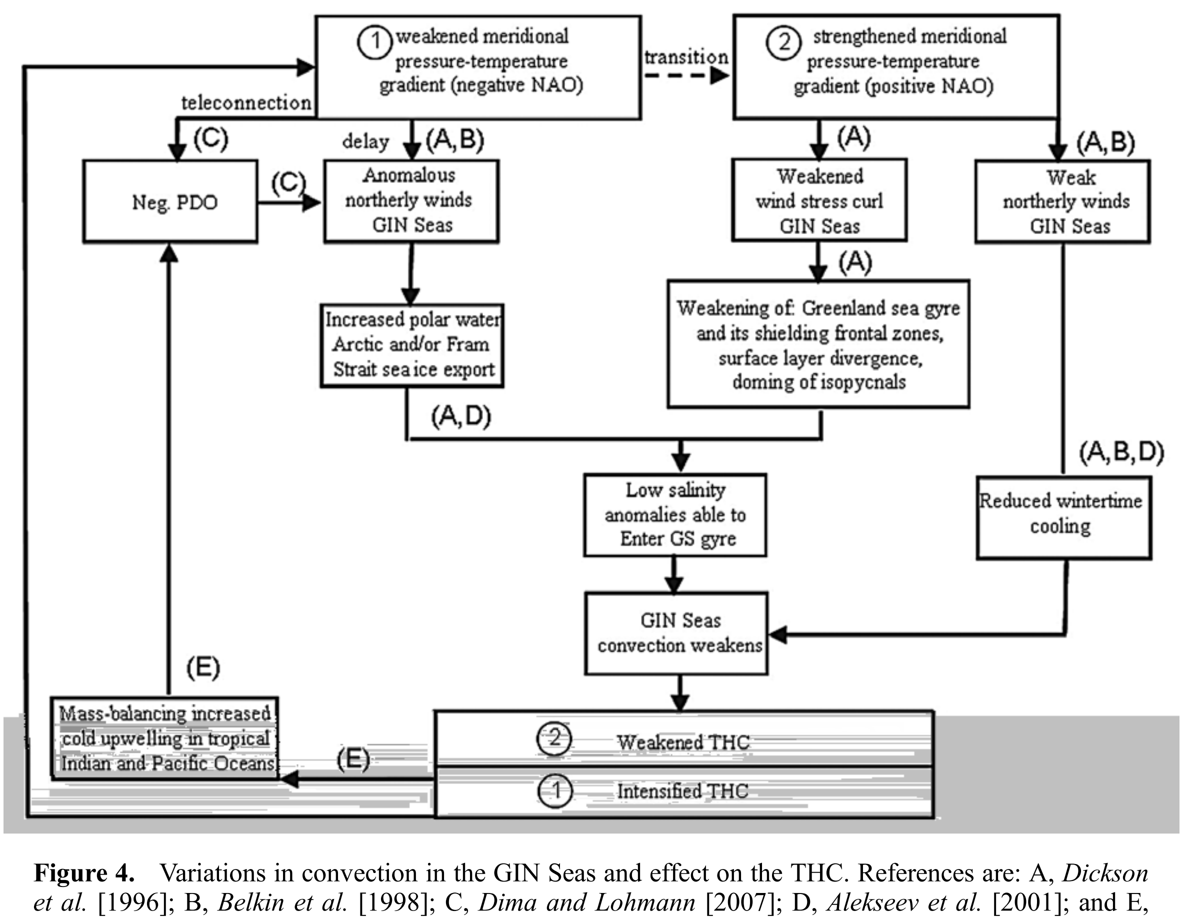

Figure 4. Variations in convection in the GIN Seas and effect on the THC. References are: A, *Dickson et al.* [1996]; B, *Belkin et al.* [1998]; C, *Dima and Lohmann* [2007]; D, *Alekseev et al.* [2001]; and E, *Gray* [1997].

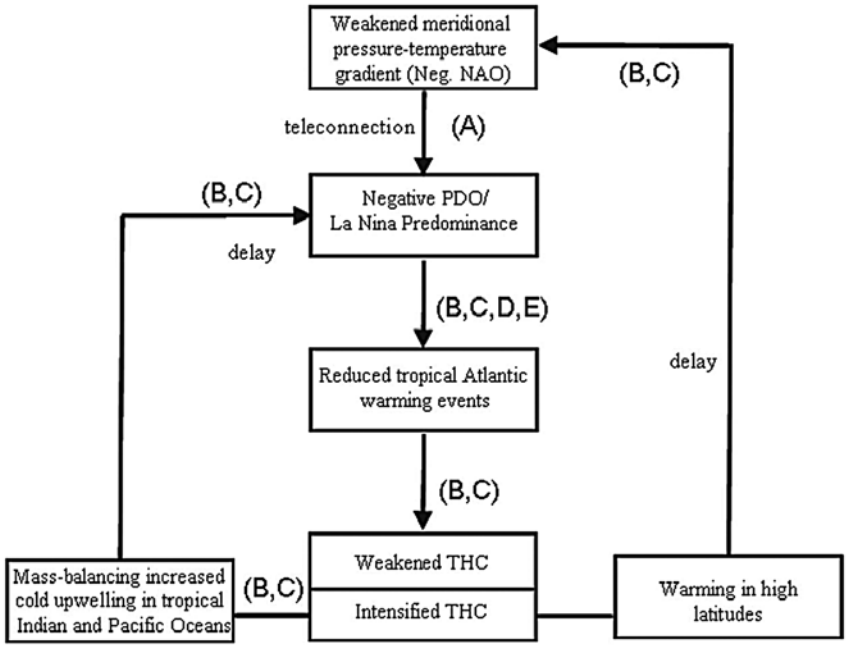

Figure 5. Effect of tropical Atlantic warming events on the THC. References are: A, *Dima and Lohmann* [2007]; B, *Gray* [1997]; C, *Gray et al.* [1997]; D, *Enfield and Mayer* [1997]; and E, *Alexander et al.* [2002].

AMO is mainly dependent on Labrador Sea convection—despite the much smaller magnitude of Labrador Sea deep water [*Dickson and Brown*, 1994], or whether the observed negative AMO phase between approximately 1970 and the mid 1990s and the subsequent positive AMO phase [*Goldenberg et al.*, 2001] may have been caused by a combination of the changes in both convective areas (i.e., weakened convection in the Labrador Sea in the 1970s, progressively weakened GIN Sea convection during the 1980s, and a strengthening of Labrador Sea convection from the mid 1980s). It is also unclear whether convection in the GIN Seas continues to be significantly reduced, thus potentially weakening the THC [*Alekseev et al.*, 2001] or whether it has now resumed. *Vinje et al.* [2002] report enhanced GIN Sea convection during winter 2001–2002. *Ronski and Budéus* [2005] document an increase in surface salinity in the GIN Seas during the 1990s and a deepening of the mixed layer to 1600 m during winter 2001–2002, compared to an average depth of 1175 m over the previous 8 winters.

3.2.2.4. Effects of Tropical Atlantic Warming Events on the THC

[35] *Gray* [1997] and *Gray et al.* [1997] propose a Northern Hemisphere Pacific-Atlantic feedback that affects the THC through variations in tropical Atlantic warming events (Figure 5). During periods of enhanced NADW formation, the mass balancing increase in cold water upwelling in the tropical Indian and Pacific Oceans should act to reduce the Pacific Warm Pool. This will initiate a shift to a negative state of the Pacific Decadal Oscillation (PDO) along with a tendency toward fewer or less intense El Niño events [*Gray*, 1997; *Gray et al.*, 1997; *Goldenberg et al.*, 2001]. A shift to a negative PDO may also result due to atmospheric teleconnections that weaken the Aleutian Low when the Icelandic Low is weakened [*Dima and Lohmann*, 2007]. A tendency toward a weakened Aleutian Low or SST patterns corresponding to a negative state of the PDO have also been reported in modeling studies [e.g., *Timmermann et al.*, 1998, Figure 21d]. Tropical East Pacific cooling and reduced occurrence of intense El Niño events is expected to lead to less frequent tropical Atlantic warming events via the atmospheric bridge mechanism [*Enfield and Mayer*, 1997; *Alexander et al.*, 2002]. This will contribute to a shift of the Atlantic atmosphere-ocean system back to a negative AMO and AMM phase (Figure 5).

3.2.2.5. Effects of Intergyre Gyre and SPG Heat Transport on the Atmosphere

[36] *Marshall et al.* [2001b] present a delayed oscillator model coupling changes in ocean gyres, the THC and Ekman transport to the atmospheric jet stream (Figure 6, left). Associated with changes in the NAO between predominantly positive and negative states are changes in the intensity and position of the STG and SPG and in the location of the high latitude positive wind stress curl and lower latitude negative wind stress curl that contribute to driving, respectively, the SPG and STG. During a positive (negative) NAO state, the zero wind stress curl line separating the two gyres has a more meridional (zonal) orientation and is shifted slightly northward (equatorward). This is associated with a strengthened (weakened) and more poleward (equatorward) positioned STG and more meridional (zonal) storm tracks. This forcing pattern is visualized as driving a circulation anomaly at the boundary between the gyres, which *Marshall et al.* [2001b] refer to as the "intergyre gyre." During the positive (negative) NAO state, the circulation of the intergyre gyre is anticyclonic (cyclonic), corresponding to a strengthened (weakened) SPG and enhanced (reduced) northward heat transport (Figure 6,

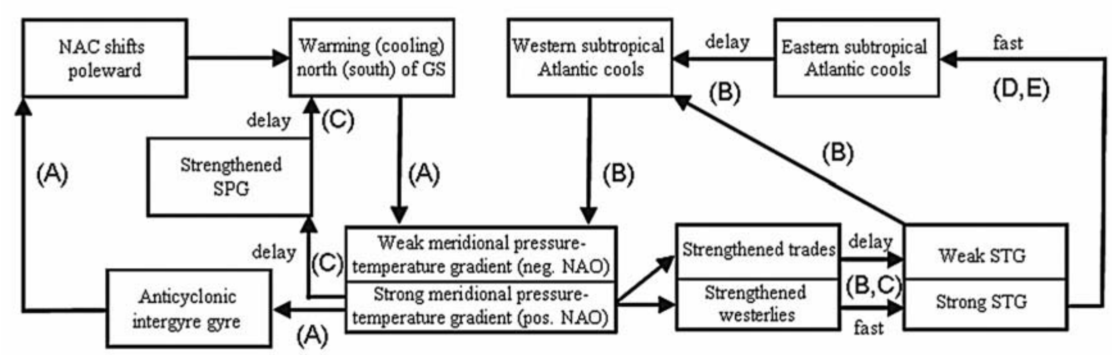

Figure 6. Effects of STG, SPG, and intergyre gyre heat transport on the atmosphere. References are: A, *Marshall et al.* [2001b]; B, *Grötzner et al.* [1998]; C, *Eden and Willebrand* [2001] and *Eden and Greatbatch* [2003]; D, *Bjerknes* [1964]; and E, *Gray* [1990].

left). A strengthened SPG with enhanced northward heat transport is also found in modeling studies after a considerable delay [*Eden and Greatbatch*, 2003] (Figure 6).

[37] The negative feedback mechanism in this framework results from the anomalous warming (cooling) to the north (south) of the zero wind stress curl line and the atmospheric jet stream that acts to reduce the meridional temperature gradient. The reduced temperature gradient weakens the jet stream and with it surface winds and pressure patterns, including the Icelandic Low (Figure 6, left). As the atmosphere transitions to a predominantly negative NAO state, the intergyre gyre changes gradually to a cyclonic flow, thus initiating the reverse part of the cycle. The role of the intergyre gyre in this framework can also be played by changes in the northward heat transport by the THC in response to the thermal anomalies generated by the NAO [*Marshall et al.*, 2001b]. The delay in this framework is provided, respectively, by the response of the intergyre gyre and the THC.

3.2.2.6. Effects of STG Heat Transport on the Atmosphere

[38] During a long-term positive NAO, more heat becomes available for northward transport along the U.S. East Coast along the western limb of the STG [*Bjerknes*, 1964]. Consequently, the STG is strengthened with a delay of a few years. At the same time, a negative anomaly develops in the eastern subtropical Atlantic due to the strengthened trade winds. In the coupled general circulation model (GCM) study presented by *Grötzner et al.* [1998], these negative anomalies propagate to the western part of the basin via the strengthened STG. The resulting reduced meridional temperature gradient feeds back on the atmospheric circulation and eventually initiates a phase reversal (Figure 6, right). A second component of this feedback results from negative wind stress curl anomalies between 30 and 50°N and positive curl anomalies south of 30°N caused by the strengthened westerlies and trades. These northern (southern) wind stress curl anomalies should act, respectively, to accelerate (decelerate) the STG. However, propagation times in the northern portion of the gyre are too long to be able to affect the STG. Only the decelerating southern curl anomalies actually modify the gyre within the necessary time frame. The resulting slowing of the STG contributes to a phase reversal since a strong STG is necessary to maintain positive SST anomalies off the U.S. East Coast (Figure 6, right). Given the delay required for the ocean and atmosphere to adjust to each other, one oscillation in this model takes about 17 years.

4. Observations of Atlantic Multidecadal Variability

[39] Following early observations of multidecadal variability in North Atlantic SLP and SST patterns by *Bjerknes* [1964], many studies have identified warm and cold phases in the North Atlantic [*Goldenberg et al.*, 2001; *Dima and Lohmann*, 2007; *Klotzbach and Gray*, 2008]. Proxy data implies that multidecadal variability in the tropical Atlantic may extend back several centuries or longer [*Delworth and Mann*, 2000; *Fischer and Mieding*, 2005]. Since long-term trends in the NAO have been discussed extensively elsewhere [e.g., *Van Loon and Rogers*, 1978], we will focus on AMO and AMM variability.

[40] *Goldenberg et al.* [2001] used far North Atlantic SST data from the mid 1870s through 2000 and established positive phases of the AMO between approximately 1875–1899, 1926–1970, and 1995–2000, and negative phases of the AMO between approximately 1900–1925 and 1971–1994 (Figure 7, top). *Klotzbach and Gray* [2008] found almost identical positive and negative phases using a combined index of far North Atlantic SSTs along with SLPs from 0 to 50°N, 70–10°W. Positive phases of the AMM tend to agree with positive phases of the AMO, as *Kossin and Vimont* [2007] have documented positive phases of the AMM between approximately the late 1870s to 1900, 1925–1970, and from 1995 to the present, with negative AMM phases being identified in the interim times.

[41] All of these assessments are based on SST and SLP measures. It would be instructive to be able to assess the strength of the THC directly through salinity and temperature (and hence density) measurements; however, extensive salinity measurements have not been accessible until recently. The development of the Simple Ocean Data Assimilation (SODA) reanalysis allows for simple calculations of salinity and temperature from 1958 to 2004 [*Carton et al.*, 2000a, 2000b]. We calculate winter averages of density based on salinity and temperature for the larger area contributing to deep water formation in the GIN Seas (64–80°N,

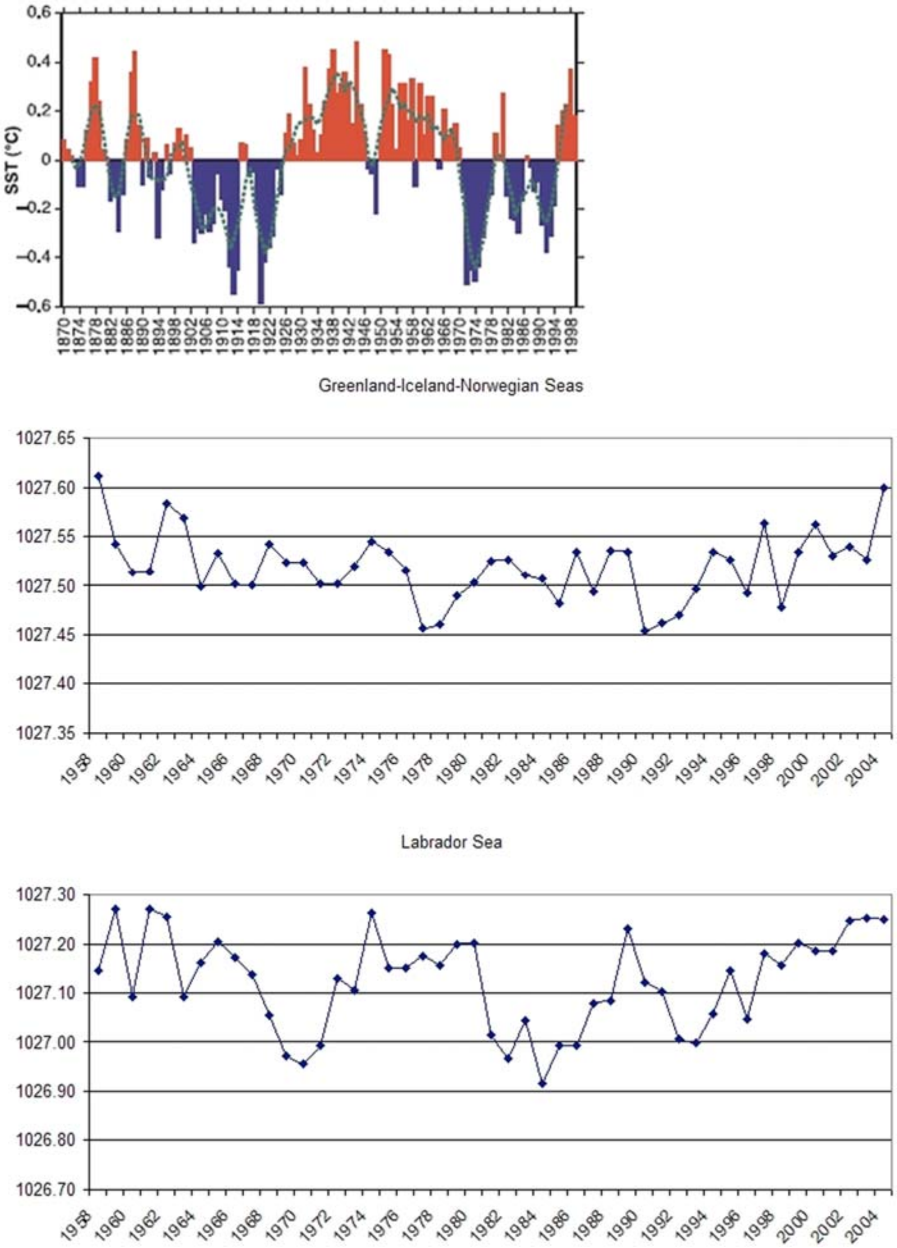

Figure 7. (top) First detrended EOF of non-ENSO related sea surface temperature variability in the far North Atlantic Ocean. Figure taken from *Goldenberg et al.* [2001]. Reprinted with permission from AAAS. (middle) Winter density averages in the GIN Seas (64–80°N, 20°E–24°W) in kg/m^3. (bottom) Winter density averages in the Labrador Sea (50°–65°N, 70°–40°W) in kg/m^3.

20°E–24°W) and the Labrador Sea (50–65°N, 70–40°W) over the period from 1958 to 2004 (Figure 7). These density calculations tend to be characterized by higher densities (and a likely enhanced THC) during the late 1950s through the 1960s, lower densities in the 1970s–early 1990s and higher densities since the mid 1990s. They do not completely match trends in the AMO, which we think is due to the role played by other factors, in particular structural properties of the

Figure 8. (a) Wind strength index in relative units for OWS Bravo (56°30′N, 51°W), from *Dickson et al.* [1996] (used with permission from Elsevier). (b) Sea ice severity near Iceland since 1900, from *Dickson et al.* [1988] (used with permission from Elsevier). The index is the product of the number of weeks with ice per year and the number of coastal areas near which it was observed. (c) (top and middle) Summer and winter salinity averages 1902–1939 in the GIN Seas (∼58°–63°N, 10°–15°W) and (bottom) summer salinity averages for the Irminger Sea (∼58°–63°N, 34°–39°W) (winter values not available), from *Smed* [1943]. (d) Salinity of North Atlantic and Arctic intermediate water in the Faroe-Shetland Channel, 1902–1982. Figure from *Dooley et al.* [1984]. Used with the kind permission of the International Council for the Exploration of the Sea.

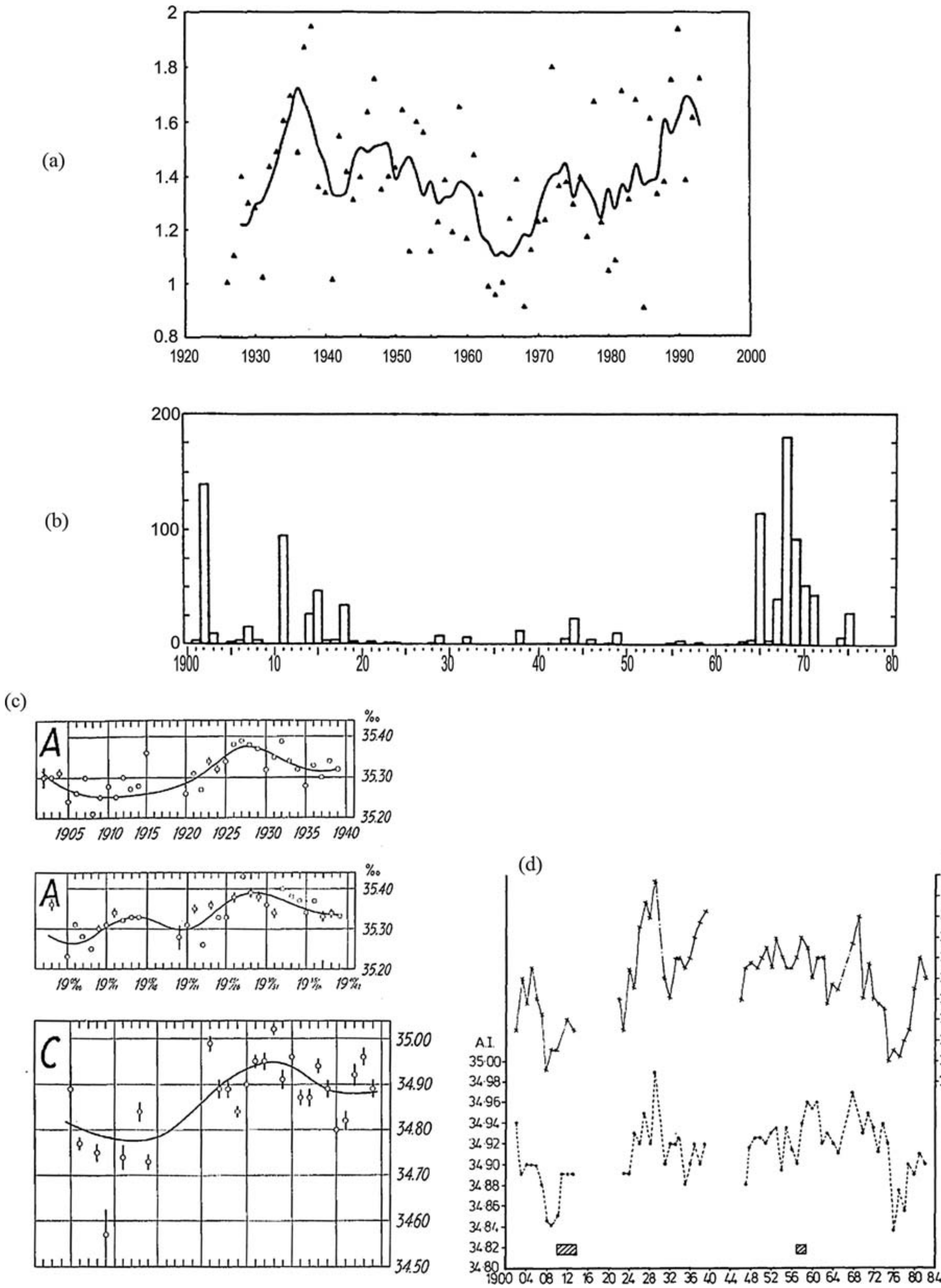

Figure 8

cyclonic gyres in the deepwater formation sites. Additional research is certainly needed in this area.

[42] Measurements of salinity values prior to 1958 have been reported, for instance, by *Smed* [1943] and *Dooley et al.* [1984]. These indicate a likely earlier salinity minimum between approximately 1905–1923 (Figures 8c and 8d). Data on sea ice severity near Iceland from 1900 as reported by *Dickson et al.* [1988] and *Wallevik and Sigurjonsson* [1998] show a maximum in the early 1900s, followed by about 15 years of higher sea ice levels (Figure 8). Arguably,

sea ice anomalies near Iceland could also be driven by local processes [*Zhang and Vallis*, 2006]. However, similar patterns are apparent in time series of Barents/Greenland sea winter sea ice anomalies [*Deser and Timlin*, 1996, Figure 2]. These higher sea ice levels may indicate a period of weakened convection due to heightened freshwater influx into the Labrador Sea. A time series of wind strength, another main contributor to variations in Labrador Sea convection, is given by *Dickson et al.* [1996]. These data indicate heightened wind stress with enhanced wintertime cooling and thus likely strengthened convection near the beginning of the positive AMO phase during the earlier part of the century, a subsequent weakening prior to the onset of the next negative phase, and a strengthening from the mid 1980s (Figure 8).

5. Summary and Future Work

[43] This paper discusses the current knowledge of three modes of natural multidecadal variability that have been identified in the North Atlantic Ocean. We discuss in detail potential mechanisms from which the AMO, the AMM and the NAO may arise and through which they interact and feed back on each other.

[44] Given these hypotheses, we examine multidecadal variability in the North Atlantic as identified in the observational record. We provide some new data from the SODA to lend credence to the hypothesized relationship between multidecadal variability in the North Atlantic and the strength of the THC.

[45] Observational evidence for multidecadal variability in the North Atlantic only extends back to the latter part of the 19th century. Since many different interrelating mechanisms have been considered in this paper, we believe that a more thorough documentation of multidecadal variability extending back to the late 1700s can be conducted utilizing the limited temperature and SLP data that are available. We intend to investigate the potential identification of active/inactive AMO phases back to the latter part of the 18th century in future research.

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