Greenland Sea Surface Temperature Change and Accompanying Changes in the Northern Hemispheric Climate

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ABSTRACT

A sudden change in the reference Greenland Sea surface temperature (GSST) in 1979 is identified. It is found to be a part of complex changes in the northern North Atlantic seas. The GSST change, in particular, resulted in a major change in the near-surface baroclinicity in the region, in addition to a large change in the net surface heat flux at the air–sea boundary over the Greenland Sea. The differences in the atmospheric mean state between two periods, one before and the other after the GSST change in the late 1970s, resemble those between the high and low North Atlantic Oscillation (NAO) index states. In addition to the changes in the mean state, major changes in the interannual variability of the atmosphere are found. A particularly interesting change in the interannual variability is found in the relationship between July GSST and the NAO phase in the following February. There is a strong correlation between July GSST and the NAO phase in the following February before the late 1970s but not at all after the late 1970s. The characteristics of these changes suggest that they may be a part of the high-frequency details of the Atlantic multidecadal oscillation.

1. Introduction

Interactions between the atmosphere and oceans in the extratropics have been studied extensively by many researchers. The North Atlantic basin has received special attention owing to the anticipated strong impacts of changes in the Gulf Stream and its downstream branches, and freshwater input into the ocean on the hemispheric and global climate. In particular, the possibility of a sudden collapse of the North Atlantic branch of the thermohaline circulation (e.g., Broecker et al. 1985) and its climatic ramifications invited intense research efforts to the study of large-scale air–sea interactions in the North Atlantic basin in the past few decades. Observational data and data products have been analyzed to identify the variability of various temporal and spatial scales in sea surface temperature (SST) and the associated atmospheric variability (e.g., Frankignoul 1985; Kushnir 1994; Czaja and Frankignoul 1999; Dima and Lohmann 2007). They have identified coupled atmosphere–ocean variability on regional to global scales and a wide range of temporal scales, ranging from interdecadal to multidecadal, connected to the North Atlantic Ocean. These diagnostic studies found that the oceanic anomalies force the atmosphere at long time scales, interdecadal and longer, and the atmospheric anomalies force back the ocean, resulting in complex loops of feedbacks at various time scales. Air–sea interactions on long time scales from interdecadal to multicentennial have also been studied using numerical models of various complexity, ranging from simple coupled box models (e.g., Nakamura et al. 1994) to coupled atmosphere–ocean general circulation models (GCMs) (e.g., Delworth and Mann 2000). Some of the coupled GCMs successfully simulate climate variability induced by low-frequency air–sea interactions (e.g., Delworth and Mann 2000), contributing to gaining insight into how the large-scale air–sea interactions may generate low-frequency climate variability. On the other hand, atmospheric GCMs forced with anomalous extratropical SST have produced confusing results, showing some atmospheric response in certain studies while showing none in other studies (Kushnir et al. 2002).

The study of the large-scale extratropical air–sea interactions has expanded in the last decade to atmospheric and oceanic processes near oceanic fronts along the two major western boundary currents, the Gulf Stream and Kuroshio/Oyashio Extensions, as these currents appear
to contribute to the maintenance of the storm tracks and general circulation in the Northern Hemisphere (Hoskins and Valdes 1990). Naturally, variations in these currents and fronts are expected to exert anomalous forcing on the atmosphere, which in turn forces back on the underlying ocean in an anomalous manner. Such interactions are bound to induce further complexity in the extratropical climate. Characteristics of atmospheric and oceanic interactions involving these oceanic fronts have been investigated and documented by a number of researchers [See Kelly et al. (2010) and Kwon et al. (2010) for comprehensive reviews on this topic]. While clear signs of atmospheric anomalies generating anomalies in the extratropical ocean and oceanic fronts at monthly to interannual time scales have been found in observational data and numerical experiments, evidence of SST anomalies along these fronts consistently generating anomalous atmospheric circulation has been difficult to find in data.

In recent works by Nakamura and Yamane (2009, 2010) and Nakamura (2012), possible impacts of SST anomalies along oceanic fronts in the midlatitudes on the large-scale atmospheric state are found in reanalysis data. They diagnosed the atmospheric reanalyses and SST data with the near-surface baroclinic vector $\mathbf{B}$ as the central parameter and found that the large-scale atmospheric anomalies can be generated by SST anomalies when the anomalies modify $\mathbf{B}$ in the vicinity of the storm tracks and upper-tropospheric jet. They defined the near-surface baroclinic vector by $\mathbf{B} = B^x \mathbf{i} + B^y \mathbf{j}$, where

$$B^x = -\frac{g}{\theta_{2m}} \frac{\partial \theta_{2m}}{\partial y},$$

and

$$B^y = -\frac{g}{\theta_{2m}} \frac{\partial \theta_{2m}}{\partial x},$$

with $\theta_{2m}$ being the monthly-mean potential temperature at 2 m above the surface, the gravitational constant denoted by $g$, and the Brunt–Väisälä frequency denoted by $N$. They demonstrated some cases in which $\mathbf{B}$ anomalies are directly related to the underlying SST anomalies and exerting substantial impacts on the large-scale atmospheric state. Their findings are consistent with recent results from numerical models that demonstrate major impacts of strong thermal fronts at the sea surface on the extratropical atmospheric circulation (e.g., Feliks et al. 2004; Taguchi et al. 2009; Sampe et al. 2010).

Though not in the core of the storm track, the Greenland Sea (GS) is located at the northern flank of the North Atlantic storm track. The GS receives very cold water of Arctic origin from the north and relatively warm water carried by the North Atlantic Current from the south. Consequently, a sharp thermal front is created where cold and warm waters meet. In addition, the land/ice surface of Greenland and the surrounding ocean generate very large surface temperature gradients during winter. One may thus wonder if anomalies in the lower boundary, ocean and sea ice, have nonnegligible impacts on the regional and/or hemispheric atmospheric state by modifying $\mathbf{B}$. Deser et al. (2000) indeed found evidence for the direct impact of sea ice in this area on the storms in the region, presumably by modification of the surface heat flux at the air–sea boundary. The impact on the storm counts may also be attributed to the change in $\mathbf{B}$. This finding by Deser et al. leads to the question: to what extent does variability in the oceanic currents in the Nordic seas affect the regional and, perhaps, the hemispheric climate on the monthly and seasonal time scales? A number of studies have, indeed, reported substantial temporal variability in oceanic currents in the Nordic seas (e.g., Adlandsvik and Loeng 1991; Flatau et al. 2003; Macrander et al. 2005; Kieke and Rhein 2006). Strong variability in the Fram Strait sea ice export (FSSIE) (Schmith and Hansen 2003) further complicates the picture of SST variations in the region.

In the following, I will present the possibility of climate variations and changes induced by changes in the GS. Section 2 describes the data and diagnostic procedures. Section 3 presents diagnoses of the atmosphere and SST based on the first modes of $B^x$ and $B^y$ in the vicinity of the GS. Section 4 describes an index defined as a measure of SST variations in the GS and its temporal characteristics. Section 5 presents and discusses differences in the climate before and after 1979. A change in the relationship between the summer GS SST and NAO in the following winter that appeared to occur in the late 1970s is described in section 6. Some discussion on the results is presented in section 7, followed by a brief summary and remarks in section 8.

### 2. Data and calculation procedures

The SST data used in this study are the Hadley Centre sea surface temperature (Rayner et al. 2003). The atmospheric data were obtained from two sources: the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40) (Uppala et al. 2005) and the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) re-analyses (Kalnay et al. 1996).

Monthly mean data of temperature at 2 m above the surface ($T_{2m}$), geopotential height ($Z$), zonal wind ($U$), meridional wind ($V$), temperature ($T$), and the net
surface heat flux ($F_h$) were obtained from the ERA-40 reanalyses. The accuracy of the $F_h$ data used here, as true for other reanalyses surface heat flux products, may not be sufficiently high to produce reliable anomaly composites. In particular, difficulty in constraining the model-calculated $F_h$ by observational data, especially during extreme events, has been noted (e.g., Josey 2001; Renfrew et al. 2002).

The activity of synoptic-scale high-frequency transient eddies is measured by $\overline{\nabla \psi}$ and $\overline{\nabla \theta}$, where $\psi$ and $\theta$ are the band-passed (period 2–7 days) $V$ and potential temperature $\theta$. An overbar denotes the ultralow-frequency (period of 30 days and longer) background field here. For these eddy quantities, simple time filters (Lau and Lau 1984) and 6-hourly fields of $T$ and $V$ from the NCEP reanalyses were used. The horizontal wind and $T$ at pressure levels from the NCEP reanalyses are very similar to those from the ERA-40 reanalyses. Also, the eddy fields calculated from the NCEP reanalyses are used only for comparison with themselves in this work. Therefore, the mixed use of NCEP and ERA-40 data in this case is not likely to affect the results and conclusions presented.

The monthly mean near-surface baroclinic vector $\mathbf{B}$ was calculated from the monthly mean ERA-40 reanalyses and NCEP surface pressure data. The data used to compute $\mathbf{B}$ are the monthly mean $T_{2m}$ and $T$ from the ERA-40 (Uppala et al. 2005) and the monthly mean surface pressure data from the NCEP–NCAR reanalyses (Kalnay et al. 1996). The ERA-40 $T_{2m}$ data rather than the NCEP–NCAR reanalysis products were chosen for its explicit inclusion of the observed near-surface temperature in producing the $T_{2m}$ data. The monthly mean surface pressure data from the NCEP–NCAR reanalyses were used to determine the pressure levels to be used for the $\mathbf{B}$ calculation and to calculate $\theta$ at 2 m above the surface from $T_{2m}$. The NCEP–NCAR surface pressure data were used for convenience because the dataset had been already compiled for calculating transient eddy fluxes used in another study and because the ERA-40 surface pressure data (not the mean sea level pressure data) at the same grid points are not readily available. The vector $\mathbf{B}$ was computed by calculating the horizontal gradient in $\theta_{2m}$, using the centered finite differencing and calculating $N$ from the lowest three vertical pressure levels above the actual atmospheric surface that are location-dependent owing to topography.

### 3. Dominant modes of $\mathbf{B}$ in the vicinity of Iceland and accompanying anomalies in the atmosphere and SST

The region surrounding the Greenland Sea is characterized by very large values in $\mathbf{B}$ and its variance (not shown). The large values are seen over most of the GS and along the eastern coast of Greenland, reflecting the presence of a very large surface temperature gradient between the relatively warm water supplied by the Gulf Stream and very cold water of Arctic origin. The boundary between the land (ice) mass of Greenland and ocean also contributes to the large values in $\mathbf{B}$ and its variance, owing to the large difference in the heat capacity between the ocean and land (ice) surface. The largest values of $\mathbf{B}$ variance during midwinter in this region are found to the north of Iceland between 20°W and 0°, where the sea ice concentration and SST show large variability (Deser et al. 2000; Furevik 2000).

To explore possible impacts of SST anomalies (SSTAs) in the GS and its vicinity on the large-scale atmospheric state, empirical orthogonal functions (EOFs) of $B^x$ and $B^y$ were calculated for each calendar month in a domain defined between 65°–75°N and 20°–3°W for the entire ERA-40 period, September 1957–August 2002. The first modes, or EOF1s, of $B^x$ and $B^y$ are found to explain more than 50% of the total variance and are separated well from the second modes, according to the criterion suggested by North et al. (1982), for most of the cold and cool months: January–April and October–December. For example, the EOF1s of $B^x$ ($B^y$) for February explain 64% (57%) of the total variance, while their EOF2s explain only about 13% of the total variance. The application of the EOF analysis to $B^x$ and $B^y$ in this small domain make it easier to isolate the patterns of $B^x$ and $B^y$ that are directly related to SST anomalies in the area. Anomalies in the SST, $T_{2m}$, $Z$, $U$, $V$, $T$, $\overline{\nabla \psi}$, $\overline{\nabla \theta}$, and $F_h$ associated with EOF1s of $B^x$ and $B^y$ are examined by forming composites for the Northern Hemisphere, using those years in which the value of the principal component (PC) of the EOF1s is $0.5 \pm 0.5$ standard deviation or larger (smaller) for the positive (negative) phase. Since February will be shown and discussed in the remainder of the paper, it is shown here also as an example. The years used for February EOF1s anomaly composites are given in Table 1. Note that the positive $B^x$ and $B^y$ have 12 years in common, while negative $B^x$ and $B^y$ have 9 years in common.

Anomaly composites of these fields for cold months suggest that anomalous $\mathbf{B}$ characterized by the $B^x$ and $B^y$ EOF1s in the vicinity of Iceland are accompanied by SSTAs in the area of large SST and ice concentration variabilities in the GS (Fig. 1). The sense of the anomalous $B^x$ is such that positive SSTAs in the positive phase of $B^x$ EOF1 in the area is associated with a northward shift in the band of large climatological $B^x$, and vice versa (Figs. 1a,c), whereas the sense of the anomalous $B^y$ is such that positive SSTAs in the positive phase of $B^y$ EOF1 in the area is associated with an enhanced $B^y$ with a slight westward shift over the GS, and vice versa (Figs.
The vicinity of the GS is on the northern flank of the North Atlantic storm track and is an area of storm propagation and generation (e.g., Deser et al. 2000). The aforementioned anomalous $B^s$ and $B^v$ in the area are, thus, expected to contribute to anomalous storm activity and large-scale circulation to some degree. Composited anomalies in $U$ and $V$ for EOF1 of $B^v$ and $B^s$, respectively, show the potential impact of the anomalous $B^v$ and $B^s$ more clearly than eddy fields. In particular, anomalous $V$ associated with $B^v$ EOF1 shows a multi-pole structure that straddles Greenland with magnitudes comparable to the local climatological values in some cases. Figure 2 shows examples of the composited anomalous $V_{200}$ associated with $B^v$ EOF1s for February and December. (A numeric subscript is used to indicate the pressure level in hPa.) Patterns very similar to composited anomalous $V_{200}$ associated with the $B^v$ EOF1 are obtained when the $B^v$ EOF1 is regressed against $V_{200}$ in most cases also. In most cases, the correlation between $B^v$ and $V_{200}$ is found to be significant somewhere in the middle and high latitudes. The areas of significant correlation generally exhibit wavelike spatial patterns (Fig. 2). Note that the amplitude and location of the multipole centers vary with case. For example, anomalous $V_{200}$ in December composites show substantially larger magnitudes over the northern North Atlantic basin than those in the February composites with their spatial structure displaced by roughly $10^\circ$ eastward compared to the pattern in the February composite (Fig. 2, note the different color scales). Also, the spatial pattern of the anomalous $V_{200}$ of the negative phase is not necessarily a mirror image of the pattern of the positive phase for the same calendar month. Though $B^v$ is usually neglected in studies of baroclinic instability, the work by Niehaus (1980) suggests a potentially important role of $B^v$ in locally enhancing the storm growth. The signals found here may well be a manifestation of the local enhancement or suppression of baroclinic waves in the region, since the flow associated with storms that grow and decay at the same spot projects onto the mean flow when averaged in time.

Examination of the PC of the $B^v$ and $B^v$ EOF1s reveals that most of the winter months of PC values greater than 0.5, and thus positive SSTAs in the area, are found after 1978, while the winter months whose $B^s$ and $B^v$ EOF1s PC values are less than −0.5 are found mostly before 1979, indicating that the reference climate in the region may have shifted from one state to another in the late 1970s. This biased distribution of the positive and negative years is seen in the PC of the $B^v$ and $B^v$ EOF1s (not shown). In the study by Deser et al. (2000), the years of positive and negative EOF1 of the Arctic winter sea ice (which accompanies positive and negative ice anomalies

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**Table 1. Years selected for anomaly composites based on the positive (upper two rows) and negative (lower two rows) phases of the February $B^v$ and $B^v$ EOF1.**

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<thead>
<tr>
<th>Parameter and phase (in February)</th>
<th>Years used</th>
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FIG. 1. Composited differences in \(B_x^r\), \(B_y^r\), SST, \(F_h\), \(T_{2m}\), and 1000-hPa wind between the positive and negative phases of (left) \(B_x^r\) and (right) \(B_y^r\) EOF1s for February. The difference is calculated by subtracting the average of those in February whose EOF1 PC value is less than or equal to 0.5 standard deviation from the average of those in February whose EOF1 PC value is greater than or equal to 1.5 standard deviation. The years used are given in Table 1. (a) Climatological \(B_x^r\) (contours) and composited difference in \(B_x^r\) color) for the February \(B_x^r\) EOF1. (b) Climatological \(B_y^r\) (contours) and composited difference in \(B_y^r\) color) for the February \(B_y^r\) EOF1. Note that the plotting scale for the composited \(B_y^r\) difference is twice larger than that used for the composited \(B_x^r\) difference. The contour interval for the climatological \(B_x^r\) and \(B_y^r\) is the same, 2 \(10^{-5}\) s\(^{-1}\). (c) Composited differences in the SST (color) and \(F_h\) (contours) for the February \(B_x^r\) EOF1. (d) As in (c), but for the February \(B_y^r\) EOF1. (e) Composited differences in \(T_{2m}\) (color) and 1000-hPa wind (m s\(^{-1}\)) for the February \(B_x^r\) EOF1. (f) As in (e), but for the February \(B_y^r\) EOF1.
(a) Feb $V_{200}$ Diff, By EOF1+–
(b) Dec $V_{200}$ Diff, By EOF1+–

Fig. 2. Composited difference in $V_{200}$ (m s$^{-1}$, color) between the positive and negative phases of the $B^\prime$ EOF1 for (a) February and (b) December from 25$^\circ$ to 90$^\circ$N. Note the different color scaling used for February and December. Correlation coefficient between the $B^\prime$ EOF1 and $V_{200}$ for February and December is shown by white contours. Contours are −0.5, −0.4, −0.3, 0.3, 0.4, and 0.5 m s$^{-1}$. The values corresponding to the statistical significance at the 90% and 95% level based on a two-tailed Student’s $t$ test are, respectively, 0.317 (−0.317) and 0.374 (−0.374).

in the GS, respectively) identified for their compositing are also, respectively, mostly before and after 1979, suggesting that the reference climate in the region may have indeed shifted in 1979. The potentially important role of the area of the GS that shows large SSTAs associated with the EOF1s of $B^\prime$ and $B^\prime$ in the regional climate through its effects on $B$ is further investigated with a simple index defined for the SST in the area.

4. Greenland Sea surface temperature index

To facilitate examining the potential role of the SST to the east of northern Greenland in climate variability and changes, an index for the SST in the area is formed. Based on the SSTAs that accompany the first EOFs of $B^\prime$ and $B^\prime$ in the vicinity of Iceland for the 45-yr period from September 1957 to August 2002, the Greenland Sea surface temperature (GSST) index (GSSTI hereafter) is defined as the area-weighted average of the SST in a small area, elongated in the southwest–northeast direction, just to the east of Greenland between 18$^\circ$ and 8$^\circ$W in the zonal direction and between 71$^\circ$ and 77$^\circ$N in the meridional direction, shown in Fig. 3a. This is the area in which values of the composited difference in the SST between positive and negative phases of the February $B^\prime$ EOF1 is approximately 1$^\circ$C or larger. Using a slightly expanded or contracted area for the definition of the GSSTI does not affect the qualitative aspect of the results presented.

The raw and 61-month running mean GSSTI shown in Fig. 3b exhibit variability at various time scales. To filter out interannual fluctuations, a 61-month running mean filter was applied to the raw monthly GSSTI. The 61-month running mean GSSTI shows a decreasing trend in the 1960s, followed by an increasing trend in the 1970s. The increasing trend in the 1970s shows a short pause in the mid 1970s, followed by another increase in the late 1970s to the near-maximum values found during the period considered. After this large increase, the value of the smoothed GSSTI fluctuates but remains substantially higher than it did before the increase in the 1970s. These trends are seen in the raw GSSTI time series as well to some degree (Fig. 3b). The time series of both raw and 61-month running mean GSSTI suggest a change in the reference value of the GSSTI in the late 1970s.

The time series of the raw GSSTI reveals an interesting jump between February and March 1979, which is the most dominant signal of the increase in the 1970s discussed above (Figs. 3c,d). The GSSTI increased by ~2$^\circ$C from February to March 1979. Even when the climatological seasonal increase between February and March is removed, it is still a 1.9$^\circ$C jump between February and March 1979 (Figs. 3c,d). Other large anomalous changes in the GSSTI between two consecutive months are only about 1$^\circ$C or less. The anomalous $F_n$ in February 1979 in the area of sudden STA increase
between February and March 1979 is very small, compared to values found elsewhere nearby, and shows no potential relationship with the GSSTI increase (not shown). Thus, if the ERA-40 $F_h$ is reasonably accurate or, if the bias in the ERA-40 $F_h$ is removed by subtracting the climatology from its monthly values, the sudden SSTA increase from February to March 1979 is not due to the thermal forcing from the atmosphere. After this jump, the GSSTI remains generally higher than its values before 1979, though fluctuations of large amplitudes are still seen after 1979. This increase in the reference values is particularly clear in winter, but is seen in summer also (Fig. 3b), indicating that the increase is likely a manifestation of changes in the oceanic currents in the region. Before 1979, the GSSTI for winter months was often at or very close to the minimum value, imposed by the freezing temperature for the seawater, whereas it mostly remained above the minimum value after 1978 (Fig. 3b). It implies that the entire GSSTI area was covered by sea ice or had a high concentration of sea ice during the winter before March 1979 and that at least a part of the GSSTI area became sea ice free during the winter after February 1979.

Spatial patterns of low-frequency SST changes that accompany the changes in the winter GSSTI over the 40-yr period from 1959 through 1998 in the surrounding seas are shown by deviations of 5-yr mean January–March (JFM) SST from the long-term climatological JFM SST (Fig. 4). A striking characteristic in the figure is the near mirror images of the SSTAs found in the 1964–68 mean and 1989–93 mean. It is very similar to the pattern of interdecadal to multidecadal SST variations found in SST data and proxy data in this region (Kushnir 1994; Delworth and Mann 2000). Also, the gradual warming trend in the GSSTI from the early 1970s to the mid 1970s is seen to be accompanied by a rapid cooling in the Irminger Sea, south of Iceland and east of Greenland (Fig. 4). The evolution of SSTAs depicted in Fig. 4 suggests a possible presence of oscillatory changes in the SST in these seas with two centers of action: one along the boundary between the cold polar water and warm North Atlantic water and the other in the Irminger Sea. Note that the sense of the anomalous net surface heat flux shown in Fig. 4 is such that the SSTAs thermally force the atmosphere in general.

5. Climate differences

a. Mean state

The climatology of various atmospheric fields and SST for two periods, September 1957–August 1978 referred to as P1 and September 1978–August 2002 referred to as P2, was computed for each calendar month and
compared. The separation into two periods, P1 and P2, is very similar to that employed in past studies on climate shifts in the region (e.g., Hilmer and Jung 2000). Since a major change in the relationship between the summer GSST and the February North Atlantic Oscillation (NAO) in the late 1970s is presented and discussed in the next section, February is shown as an example. Figures 5 and 6 show climatology differences in some of the standard climate parameters: ΔSST, ΔT2m, ΔFh, ΔZ200, ΔU200, ΔV200, ΔθS850, and Δθ200. Here, the difference in the climatology of a variable or index Q between P1 and P2 will be denoted by ΔQ = QP2 − QP1.

There are four interesting features in ΔSST (Fig. 5a). The first one, which is the main subject of this study, is the warming in a meridionally elongated small area along the eastern coast of Greenland. This feature is seen throughout the year. Warming is also seen in a band along the Gulf Stream, extending northeastward from the eastern coast of North America, most of the year as well. The third interesting feature is the major cooling in the Okhotsk Sea found in January–April. The fourth feature to note is the warming in the East China Sea to the west and southwest of Japan most of the year. These features in ΔSST, except for that along the Gulf Stream, are accompanied by statistically significant ΔFh (Fig. 5c).

In particular, the ΔFh in the GSSTI area and Okhotsk Sea is very large in cold months, showing values in the range from 50 to 250 W m⁻², which is comparable to the difference between the high and low ice phase composites reported by Deser et al. (2000). The ΔFh in the East China Sea and along the Gulf Stream is modest, ranging from 5 to 50 W m⁻². The sense of the ΔFh in

Fig. 4. Time series of 5-yr mean SST (°C, color) and Fh (W m⁻²; contours, interval 25 W m⁻²) for January–March (JFM) in the Nordic seas. Shown is the 5-yr mean minus the long-term (1958–2002) mean. (a) 1959–63, (b) 1964–68, (c) 1969–73, (d) 1974–78, (e) 1979–83, (f) 1984–88, (g) 1989–93, and (h) 1994–98.
these areas is such that the ΔSST thermally forces the atmosphere, having the same sign as that of the corresponding ΔSST. All of these features, except for the modest ΔFh along the Gulf Stream, are statistically significant at the 95% level or higher. One note of caution is the overestimated degree of freedom for ΔSST in calculating its statistical significance. There are some areas in which the monthly SST exhibits moderately strong autocorrelation within P1 and P2. In regard to the four features of ΔSST mentioned above, the autocorrelation of the monthly values is weak and is not an issue.

The accompanying ΔT2m is large over the GS and Okhotsk Sea and is moderately large in large areas in the Northern Hemisphere (Fig. 5b). Its overall spatial structure resembles that of the anomalous T2m found to
accompany the NAO. The accompanying tropospheric and lower-stratospheric $\Delta Z$ that exhibits an equivalent barotropic structure (an example is shown for the 200-hPa level in Fig. 5d) also resembles the difference between high and low NAO states (e.g., Hurrell and Van Loon 1997; Hurrell and Deser 2009). In terms of $\Delta U$, there is a zonally elongated dipole throughout the troposphere and lower and mid stratosphere over the North Atlantic (Fig. 6a). The dipole implies that the jet over the North Atlantic shifted poleward in the late 1970s. This shift is also accompanied by a similar shift in the storm track (e.g., $\Delta \overline{\vec{u} \vec{\theta}}$ and $\Delta \overline{\vec{v} \vec{\theta}}$ shown in Figs. 6c,d). The differences in the eddy fields are similar to the difference between the high and low NAO composites of storm counts and eddy fields (Hurrell and Van Loon 1997; Deser et al. 2000; Hurrell and Deser 2009). The differences in these fields for other cold months are similar to those shown.

The $\Delta V$ shows a wavelike pattern of moderate values with respect to the local climatological values and standard deviation (Fig. 6b). Unlike $\Delta U$, which shows more or less the same spatial structure in all cold

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**Fig. 6.** As in Fig. 5, but for (a) $\Delta U_{200}$ (m s$^{-1}$), (b) $\Delta V_{200}$ (m s$^{-1}$), (c) $\Delta \overline{\vec{u} \vec{\theta}}$ (K m s$^{-1}$), and (d) $\Delta \overline{\vec{v} \vec{\theta}}$ (m$^2$ s$^{-2}$).
months, the spatial pattern and area of statistical significance of $\Delta V$ vary visibly with calendar month. The January and March $\Delta V_{200}$ are shown for comparison with February $\Delta V_{200}$ (Fig. 7). The $\Delta V$ is most likely caused by, at least partially, very large $\Delta B'$ in the vicinity of the GS associated with $\Delta SST$ and $\Delta T_{2m}$ in the GS. The large $\Delta B'$ in the vicinity of the GS exerts a meridional steering effect on the mid- and upper-tropospheric wind, thereby acting as a factor that drives the meridional meandering of the tropospheric jet. Since these meanders are prone to be transient and do not necessarily appear robust when averaged in time, $\Delta V$ induced by $\Delta B'$ does not necessarily show up clearly in the climatological difference plots. It is also possible for meanders induced by $\Delta B'$ in the vicinity of the GS to propagate horizontally and appear as significant $\Delta V$ elsewhere, depending on the mean flow and other sources of wave forcing. The regional factors arising from $\Delta B'$ in the vicinity of the GS are superimposed on the planetary-scale zonally inhomogeneous circulation differences (see Figs. 5a and 6b) that can cause $\Delta V$ themselves. These may be the reason for the noticeable differences in $\Delta V_{200}$ among cold months that have very similar $\Delta B'$. Also, the very large $\Delta B'$ in the region is accompanied by large difference in the interannual variability in the monthly $V$ in the region, as discussed in the following subsection, suggesting the possibility of a substantial nonlinear impact of $\Delta B'$ on the regional and hemispheric climate.

The climatological difference is shown and discussed for the cold months above but is present in the warmer months as well. Their magnitudes in the warmer months are, however, generally smaller than those shown above. Also, the temporal autocorrelation of the atmospheric fields shown above, except for $T_{2m}$ to the south of Iceland and east of Greenland, is very weak and does not pose a problem in evaluating the statistical significance of the climatological difference in these fields.

### b. Variability

The interannual variability, defined here as the standard deviation $\sigma$ of the monthly mean field about its climatology for P1 and P2, differs substantially between P1 and P2. How the variability differs between P1 and P2 varies visibly with calendar month. Examples of differing interannual variability are shown for February in Fig. 8 that shows $\sigma_{U_{200}}$, $\sigma_{V_{200}}$, $\sigma_{Z_{200}}$, and $\sigma_{T_{2m}}$ for P1 and P2, and their differences between P1 and P2, $\Delta \sigma_{U_{200}}$, $\Delta \sigma_{V_{200}}$, and $\Delta \sigma_{T_{2m}}$. The $\Delta \sigma_{U}$ over the North Atlantic basin reflects the $\Delta U$. It is larger over northern Europe and smaller in the central North Atlantic basin and southern Europe in P2 than in P1. Thus, a simple picture of a change in variability occurring as a result of a shift in the jet axis is found for the North Atlantic basin. The normalized difference in the $\sigma_{U_{200}}$ and $\Delta \sigma_{U_{200}}/0.5[\sigma_{U_{200}}(P1) + \sigma_{U_{200}}(P2)]$ exceeds 50% over western Europe (not shown). For $V$, an even larger variability difference is found between P1 and P2, particularly clearly in the northern North Atlantic basin with the maximum difference found about where large $\Delta B'$ is generated as a result of $\Delta SST$ in the area of GSSTI. The $\sigma_{V_{200}}$ is substantially reduced, by 80% or
Fig. 8. Standard deviation of various monthly mean fields, between 25° and 90°N, for (left) P1 and (center) P2; (right) difference between P1 and P2 (P2 – P1): (from top to bottom) $U_{200}$ (m s$^{-1}$), $V_{200}$ (m s$^{-1}$), $Z_{200}$ (m), and $T_{2m}$ (K).
more locally when measured by $\Delta \sigma_{V_{20}}/0.5[\sigma_{V_{20}}(P1) + \sigma_{V_{20}}(P2)]$, in P2 compared to P1 over the GS and northwestern North Atlantic basin (not shown). This conspicuous reduction in $\sigma_{V_{20}}$ in P2 over the GS and northwestern North Atlantic basin is most likely a direct ramification of the reduced sea ice concentration and/or coverage in the region. This hypothesis is based on the reduced $\sigma_{T_{2m}}$, which is directly related to the reduced $\sigma_{2m}$ (Fig. 8i) through reduced $\sigma_{B_l}$ (not shown), while $\sigma_{SST}$ is increased (not shown). Since $\sigma_{T_{2m}}$ and $\sigma_{B_l}$ are substantially larger in P1 than in P2, while the underlying $\sigma_{SST}$ is substantially smaller, one can deduce that the effects of sea ice in the GSSTI area are responsible for the greater $\sigma_{2m}$, $\sigma_{B_l}$, and $\sigma_{V_{20}}$ in P1. The $\Delta \sigma_{U}$ and $\Delta \sigma_{V}$ are accompanied by substantial $\Delta \sigma_{Z}$ and $\Delta \sigma_{T_{2m}}$. The $\sigma_{Z}$ in P1 is visibly larger over northern North America, the North Atlantic, and northern Europe than in P2, while it is substantially smaller over the North Pacific than in P2 (Fig. 8i). As anticipated from the reduction in $\sigma_{U}$ (Fig. 8f), which is closely related to the poleward atmospheric heat transport variability, the $\sigma_{T_{2m}}$ is reduced significantly, exceeding 80% locally when measured by $\Delta \sigma_{T_{2m}}/0.5[\sigma_{T_{2m}}(P1) + \sigma_{T_{2m}}(P2)]$, in the vicinity of the area of GSSTI (Fig. 8f). On the other hand, it is substantially increased over the Okhotsk Sea where the reference winter SST decreased in the late 1970s (Fig. 8f). In addition to these areas, a major portion of North America, Eurasia, and the Arctic region exhibits substantial changes in the $\sigma_{T_{2m}}$ between P1 and P2 (Fig. 8f). These changes in variability between P1 and P2 are found, though with slight quantitative changes, even when the variability for P1 and P2 are calculated against the climatology for the entire 45 years, September 1957–August 2002.

### 6. GSST and NAO

The relationship between the NAO and the Nordic seas has been studied by a number of researchers in the past (e.g., Cayan 1992; Deser et al. 2000; Dickson et al. 2000; Saloranta and Haugan 2001; Flatau et al. 2003). These studies found, in general, that the ocean in the region responds to anomalous atmospheric thermal and momentum forcings associated with the NAO on monthly to seasonal time scales, though substantial variability in the oceanic state in the region seems to arise from the oceanic internal dynamics as well (Saloranta and Haugan 2001; Kieke and Rhein 2006). To investigate a potential connection between the anomalous SST in the GS and the large-scale atmospheric circulation, and its impact on the monthly to seasonal climate variability in the Northern Hemisphere, the correlation between the GSSTI and NAO index (NAOI hereafter) was examined on a monthly basis with leads and lags up to 12 months. For this purpose, the NAOI, based on the surface pressure difference between Lisbon, Portugal, and Stykkishólmur, Iceland, obtained from the Hurrell NAOI data site hosted by the NCAR, was used.

The correlation between the GSSTI and NAOI is generally weak and insignificant. One exception is found, however, when the GSSTI in mid summer leads the NAOI in February in P1. The correlation is especially high and significant between July GSSTI and the following February NAOI with a correlation coefficient value of 0.712. The time series of the normalized anomalous July GSSTI and February NAOI for P1 show this relationship (Fig. 9). This strong relationship between the July GSSTI and NAOI with a 7-month lag is found only in P1 and not at all in P2. Note that the monthly GSSTI anomaly fluctuates rapidly, making its temporal autocorrelation within P1 and P2 very weak beyond one-month lag or lead. Thus, the relationship between July GSSTI and the following February NAOI is likely to be meaningful. Changes in the relationship between the NAO and North Atlantic SST, that is, between the NAO and the surface atmospheric temperature and between the NAO and the sea level pressure, have been reported by Polyakova et al. (2006). Also, Schmith and Hansen (2003) reported a strong dependence of the correlation between the NAO and FSSIE for the period analyzed. These results suggest that the North Atlantic climate variability may have, indeed, changed in various ways in the late 1970s.

The significant correlation between the July GSSTI and February NAOI is very intriguing and suggests the possibility of an extended forecast for February when the reference climate is in a state that resembles that of P1. The correlation coefficient is reduced to 0.358 when the GSSTI leads the February NAOI by six months and
disappears when the GSSTI leads the February NAOI by five months. This lagged correlation between the GSSTI and NAOI is reflected clearly in anomaly composites of various fields for the February following a July of strong signals in the GSSTI (February–July GSSTI for short). Composited anomalous $U_{200}$, $V_{200}$, $T_{2m}$, and SST for February–July GSSTI are shown in Fig. 10. The anomalies shown here are the average of the top five positive February–July GSSTI subtracted from the average of the top five negative February–July GSSTI. The years of February–July GSSTI (thus, years after the July of strong GSSTI signals) used for the composites are given in Table 2. The composited anomalies shown in Fig. 10 have spatial structures very similar to those
TABLE 2. February years selected for anomaly composites based on the positive (upper row) and negative (lower row) phases of the preceding July GSSTI.

<table>
<thead>
<tr>
<th>Parameter and phase</th>
<th>February years</th>
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associated with the NAO, though their magnitudes are somewhat smaller than those found in composites based on the February NAOI formed with the five strongest positive February NAOI years (1959, 1967, 1973, 1974, and 1976) and five strongest negative February NAOI years (1960, 1965, 1966, 1969, and 1978). The anomalous tropospheric circulation has an equivalent barotropic structure in general (not shown). The most striking feature of the anomalous circulation is the anomalous $V$ that has large (with respect to its local climatology and $\sigma$) values over the northern North Atlantic sector with northward (southward) anomalous flow in the positive (negative) phase over the Greenland and Norwegian Seas (Fig. 10b). This large positive anomalous $V$ is clearly related to very large positive anomalous $B\sigma$ over the GS (not shown) and is likely to be a manifestation of enhanced growth of storms over these seas when the SST is anomalously high and, thus, $B\sigma$ is also anomalously high. In the negative phase, the large negative $V$ is likely to be a manifestation of reduced growth of storms there and reduced $B\sigma$. When storms grow and decay in specific areas repeatedly, that part of their wind, which is not cancelled out in time averaging, shows up as a part of the mean flow. When there is an area of large baroclinicity upstream of an area of small baroclinicity, for example, storms grow and decay in a spatially inhomogeneous manner and contribute to the mean flow. In the case of an area of large positive $B\sigma$ with a spatial scale of the local Rossby deformation radius or larger, repeated localized northward steering of the atmosphere is anticipated to result in the northward mean flow. The northern fringe of this area of large anomalous $B\sigma$ is the narrow area between Greenland and the anomalously warm band in the GS. Thus, one may argue that SSTAs in the area of GSSTI are indeed responsible, at least partially, for the anomalous general circulation portrayed in Fig. 10. This hypothesis is consistent with the difference in storm counts in the area between the high and low sea ice cases examined by Deser et al. (2000).

Time series of composited anomalous SST and $F_h$, based on the July GSSTI and February NAOI, were examined and compared to study the nature of the aforementioned lagged correlation and the patterns and origins of SSTAs involved in the correlation. Figures 11 and 12 show the time series of composited anomalous SST and $F_h$ for the northern seas in the case of large July GSSTI. The anomalies are shown as the difference between the averages of the five samples each of the strongest positive and negative July GSSTI cases. The month that leads July by one is denoted $F_{h-1}$, while the month that lags July by one is denoted $F_{h+1}$, and so on. The years of January–March used for the composites are listed in Table 2. The years of June–December used are those listed in Table 2 minus one year. Time series of composited anomalous SST and $F_h$ for the case of strong February NAOI signals were also formed for a 10-month period from the June that precedes the February of strong signals in the NAOI through the March that follows the the February of strong signals in the NAOI.

The time series based on the two different indices have similar spatial structures in the anomalous SST and, to a lesser degree, in $F_h$, with the exception of $F_h$ in January. Because of this similarity, only the composites based on the July GSSTI are shown. These time series show how the correlation between the February NAOI and GSSTI disappears when the latter leads the former by five months or less. When the July GSSTI is large and positive, the SSTAs are large and positive in the area of the GSSTI from June to August (Figs. 11a,b,c) but reduce to insignificant values by September (Fig. 11d) and remain so through February. This is why the correlation between the GSSTI and February NAOI is strong only when the GSSTI leads the February NAOI by seven months or so. Note that SSTAs grow to substantial positive values in the northern Norwegian Sea in the mid winter and that SSTAs to the south of Greenland and Iceland remain negative throughout the eight-month period up to February. It may be this southwest–northeast contrast between negative and positive SSTAs in the region that plays a significant role in the positive NAOI in February during P1 (Fig. 11i).

The time series of anomalous $F_h$ in Fig. 12 depict very complicated pictures of air–sea interactions that may be involved in the relationship between the February NAOI and July GSSTI. One clear signal found in the time series based on both the July GSSTI and February NAOI is the anomalous input of $F_h$ in June and July in most of the northern North Atlantic seas (Figs. 11a,b and 12a,b). In these June and July composites, the anomalies to the north and east of Iceland are such that anomalous atmospheric forcing helped generate SSTAs since the anomalous SST and $F_h$ have opposite signs. On the other hand, to the south and west of Iceland, oceanic anomalies helped generate anomalies in the atmosphere since anomalous SST and $F_h$ have the same sign. From August through January the patterns of anomalous thermal forcing do not suggest any clear-cut picture. In
February, however, the anomalous SST and $F_h$ in the northern North Atlantic seas suggest anomalous thermal forcing of the ocean by the atmosphere for the most part, with some exceptions in a narrow band of positive SSTAs found in the vicinity of the boundary between warmer water of Atlantic origin and colder water of Arctic origin in the positive phase of February NAOI, and vice versa (Figs. 11i and 12i). The reason for the

![Diagram](image-url)

**Fig. 11.** Time series of composited anomalous SST (°C) from June that precedes strong signals in July GSSTI to the following March: (a) Mo−1, June that precedes strong signals in July GSSTI; (b) Mo0, July of strong signals in GSSTI; (c) Mo+1, August that follows strong signals in July GSSTI; (d) Mo+2, September that follows strong signals in July GSSTI; (e) Mo+3, October that follows strong signals in July GSSTI; (f) Mo+4, November that follows strong signals in July GSSTI; (g) Mo+5, December that follows strong signals in July GSSTI; (h) Mo+6, January that follows strong signals in July GSSTI; (i) Mo+7, February that follows strong signals in July GSSTI; and (j) Mo+8, March that follows strong signals in July GSSTI. The anomalies are shown by the average of five top positive years minus five top negative years; the years of January–March used for the composites are given in Table 2. Composited time series based on February NAOI have spatial structures very similar to those shown with slightly smaller magnitudes.
positive/negative GSSTI in July preceding positive/negative NAOI in the following February during PI is unclear. Nevertheless, it seems reasonable to say that a part of the initial trigger is anomalous atmospheric thermal forcing of the GS and its surrounding seas in June and July. Of course, the anomalous atmospheric heat input into the GS and its surrounding seas in June and July may be partially a product of SSTAs and/or sea ice anomalies in the region, or elsewhere, in the first place. These features are likely to be products of complex feedbacks operating in the climate system. This relationship is not visible when the seasonal mean GSSTI and NAOI are used for the diagnoses.

7. Discussion

One may wonder if the sudden and large increase in the GSSTI between February and March 1979 is an
artifact of changes in data collection methods and procedures, such as the change in the satellite sensor and passive microwave retrieval algorithm. However, SST data based on weekly maps compiled by the Norwegian Meteorological Institute, presented by Grotefendt et al. (1998, their Fig. 8c shows the maximum Fram Strait SST for February–March average sharply increasing from 1979 to 1980), John Walsh sea ice extent data presented by Mysak et al. (1990, their Fig. 3–1 shows the 3-month running mean anomalies in the sea ice extent for the northern GS sharply decreasing from 1978 to 1980), and Fram Strait sea ice export (FSSIE) presented by Schmith and Hansen (2003, their Figs. 6 and 8 show a sharp decrease in the FSSIE in the late 1970s) support the local warming event in 1979 found in the Hadley Centre SST data. It is also consistent with the findings of reduced sea ice in the Arctic region in the 1980s and 1990s by Deser et al. (2000). The cause of this warming is not clear but may be attributed to a change in position of the boundary between cold water of Arctic origin and relatively warm water of Atlantic origin, as the northward transport of warm water into this basin increased in the early and mid 1970s (Adlandsvik and Loeng 1991). Hydrographic data also suggest increased northward transport of the Atlantic water into the northern GS in the late 1970s (Saloranta and Haugan 2001). As noted in the introduction, currents in the Nordic seas exhibit strong variability and may well be responsible for this change. Because of the high variability of oceanic currents in the region and weak stratification and strong mixing in the ocean expected in the area during winter, it is difficult to conceive a $2^\circ$C SST increase in the area of GSSTI attributed to the direct atmospheric thermal forcing in February and March.

Although the discussion above is focused on the GSST and its potential role in regional and hemispheric climate, the change in the GSSTI may simply be a facet of a change in the entire Arctic and sub-Arctic region. This speculation is based on a precipitous drop in the winter reference SST in the marginal Okhotsk Sea that occurred between December 1978 and January 1979 (Fig. 13), in addition to the hemispheric scale of climatology differences shown in section 5. When the area-weighted monthly-mean SST is calculated as an index for the marginal Okhotsk Sea (OSSTI hereafter) for the area that shows large $\Delta$SST in the region (Fig. 4a), its time series shows a sudden change in its reference value only two months before the change in the reference GSSTI (Figs. 13b,c,d). This change in the reference OSSTI is found clearly only in January–March (Fig. 13c), suggesting that it is likely related to the sea ice coverage and/or concentration in the region forced by the atmosphere and is not necessarily a manifestation of shifts in

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**FIG. 13.** (a) Area used to calculate the OSSTI. (b) Time series of the raw monthly OSSTI (dotted lines with open circles at data points) and 61-month running-mean OSSTI (solid lines with filled circles at data points) for the period from 1958 to 2002. (c) Time series of the monthly OSSTI minus the 1958–2002 climatological monthly value from 1974 to 1983. Only 10 years around 1979 are shown so as to allow easier identification of the large and sudden increase between December 1978 and January 1979. (d) Time series of anomalous change in the monthly OSSTI between two consecutive months from 1974 to 1983. The anomalous change is calculated by subtracting anomalous (with respect to the 1958–2002 monthly climatology) OSSTI for the month being evaluated from anomalous OSSTI for the next month.
oceanic currents. Despite the lack of any correlation between the GSSTI and OSSTI, with or without time lags or leads, the nearly simultaneous changes in their reference values, though of opposite signs, strongly indicate that the changes are related.

The two-month lead signal in the onset of the major change found in the OSSTI and the lack of any precursor signal in the OSSTI suggests the possibility of a climate change being initiated by atmospheric anomalies on a relatively short time scale. On the other hand, a close examination of GSSTI shows the beginning of the rising trend in the early 1970s, as indicated by SSTAs shown in Fig. 4, eventually culminating in the sudden jump in 1979. Thus, it is also possible that the polar and subpolar regions went through a period of transition over several years preceding the major shift, with the oceans and ice in the northern North Atlantic playing important roles. This hypothesis is supported by a period of high winter NAOI in the early to mid 1970s and increased poleward transport of heat by the oceanic currents driven, presumably, by the high-NAOI atmospheric circulation in the early 1970s (Dickson et al. 2000). Since an increased poleward transport of warm water into the Nordic seas itself helps increase $B^P$ to the east of Greenland and favors an anomalous atmospheric circulation that resembles the positive phase of NAO, as suggested above and is predicted by theories (Niehaus 1980; Hoskins and Valdes 1990; Rhines and Schopp 1991), there is a positive feedback between the poleward oceanic heat transport into the Nordic seas and phase of the NAO. This feedback adds more complexity to the atmosphere–ocean–ice system that is very complex even when it is much simplified (Nakamura et al. 1994; Nakamura 1996; Jayne and Marotzke 1999).

The climate change presented here may also be considered as a generalized view of a zonal shift in the centers of action of the NAO in the late 1970s (Hilmer and Jung 2000; Jung and Hilmer 2001; Jung et al. 2003) that appears to be a manifestation of an increase in the frequency of occurrence of the positive phase of the NAO after the late 1970s (Cassou et al. 2004). The climate change presented here is consistent with the increased frequency of occurrence of the positive phase of the NAO after the 1970s, which is accompanied by a change in the reference state and variations about the reference state in a traditional view of the climate and its variations. In this sense, an alternative view of a climate change may be a change in the frequency of occurrence of dominant climate regimes, as suggested by the results of Cassou et al. (2004). Regardless of how one may view the climate change in the late 1970s, I hypothesize that the underlying direct cause of the change is the change in the GS resulting in a major change in the $B^P$ climatology and variations in the vicinity of the GS. Numerical experiments with high-resolution atmospheric GCMs may prove useful to examine the hypothesis.

What brought about the change in the Greenland Sea—that is, the root cause of the climate change in the late 1970s—is unclear and is beyond the scope of this work. One may be able to gain some insight into the cause of the changes in the GS shown here by viewing them as subintercentennial high-frequency details of the Atlantic multidecadal oscillation (AMO), studied by a number of researchers (e.g., Enfield and Mestas-Nunez 1999; Dima and Lohmann 2007, 2011). The sudden increase in the GSSTI between February and March 1979 is likely to be connected to the major reduction in the FSSIE in the late 1970s (Schmith and Hansen 2003). Dima and Lohmann (2007, 2011) present the reduction in the FSSIE as a large subintercentennial signal in the late 1970s. Dima and Lohmann note a close connection between the AMO and FSSIE with the latter leading the former by 10–15 years by affecting the convective activity in the regions of North Atlantic Deep Water formation. They further speculate on a possible loop of processes that forms a negative feedback and may generate an oscillatory cycle of the AMO. A critical component of this loop is the FSSIE that appears to be an important factor in the sudden change in the GS in 1979 and also in the increasing AMO index in the 1980s and 1990s. While the AMO is, by definition, used to describe multidecadal changes in the North Atlantic basin, the actual change that characterizes a change between multidecades may occur within a short period of time, say, one month. The sudden increase in the GSSTI and accompanying change in $B^P$ between February and March 1979 may be such a change. Temporal characteristics of the 61-month running mean GSSTI and the time series of the AMO index shown by Dima and Lohmann (2007) are very similar to each other with an approximate lag of 15 yr and are consistent with this interpretation. The sudden increase in the GSSTI occurring in 1979 when the effect of the Great Salinity Anomaly in the northern North Atlantic Ocean had subsided is probably not a coincidence since the warmer Nordic seas, assuming the surface temperature to its west remain the same or decrease, support a condition favorable for enhanced northward atmospheric flow through increased $B^P$ and, thus, enhanced northward oceanic heat and salt transports into the Nordic seas and a suppressed FSSIE (Dickson et al. 2000). The atmospheric surface pressure anomalies associated with the AMO presented by Dima and Lohmann (2007) show large values of opposite signs in the mid- and high-latitude North Atlantic and North Pacific basins simultaneously. Therefore, viewing the sudden changes in the
GSSTI and OSSTI in 1979, the climate change presented above, and a climate shift in the North Pacific basin in the mid 1970s reported in past studies (e.g., Nitta and Yamada 1989; Trenberth and Hurrell 1994) as parts of the AMO cycle also seems reasonable.

8. Concluding remarks

A climate change possibly resulting from a change in the reference SST and sea ice in the Greenland Sea in the late 1970s is presented and discussed. The GSST increased very sharply between February and March 1979, resulting in a change in the reference SST in the region. The accompanying difference in the mean state between two periods, one from 1957 through 1978 and the other from 1979 through 2002, resembles the difference between the positive and negative phases of the NAO. The evolution of the GSSTI and the SST in the Nordic seas suggests that the sudden change in the GSSTI and, possibly, the climate in March 1979 may represent the final “slip” into a new climate after the Nordic seas went through a gradual warming in the early and mid 1970s. The climate change in the late 1970s is found in not only the mean state but also in the interannual variability. A particularly intriguing facet of the climate change is a complete change in the relationship between the summer GSSTI and the following February NAOI. Before the change, the July GSSTI and the following February NAOI were strongly correlated. After the change, the correlation completely disappeared. The discovery of the change in relationship between the July GSSTI and February NAO adds one more piece of information to the complex picture of climate variations in the North Atlantic basin.

The mechanisms behind the sudden changes in the GSSTI and OSSTI in 1979 need to be examined carefully with a numerical model that can successfully reproduce such behavior of the system, including the preceding changes in the North Atlantic Ocean and Nordic seas and the change in relationship between the GSSTI and NAOI. It is important to address if such a change can occur without the preconditioning of the Nordic seas by sustained anomalous atmospheric forcing, as suggested above, since, if it can, it implies a possibility of a sudden climate change on the hemispheric scale occurring as a result of internal changes in the SST and sea ice in a small area of the Nordic seas. It is also important to examine if these changes are parts of a multidecadal and/or multicentennial natural cycle such as the Atlantic multidecadal oscillation. By addressing these questions, one would hope to gain further insight into the role that the greenhouse gas increase might have played in the climate changes in the last century. The results presented here demonstrate the difficulty in predicting short- to midterm climate variations, not to mention longer-term climate variations or changes, without a reasonably good representation of various processes in the Arctic and sub-Arctic regions in a climate simulation model. In this regard, the assessment of the accuracy of such processes in climate simulation models seems imperative.

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