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Interdecadal climate variability in the subpolar North Atlantic

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Abstract. The statistical relationships between various components of the subpolar North Atlantic air-sea-ice climate system are reexamined in order to investigate potential processes involved in interdecadal climate variability. It is found that sea surface temperature anomalies concentrated in the Labrador Sea region have a strong impact upon atmospheric sea level pressure anomalies over Greenland, which in turn influence the transport of freshwater and ice anomalies out of the Arctic Ocean, via Fram Strait. These freshwater and ice anomalies are advected around the subpolar gyre into the Labrador Sea affecting convection and the formation of Labrador Sea Water. This has an impact upon the transport of North Atlantic Current water into the subpolar gyre and thus, also upon sea surface temperatures in the region. An interdecadal negative feedback loop is therefore proposed as an internal source of climate variability within the subpolar North Atlantic. Through the lags associated with the correlations between different climatic components, observed horizontal advection time scales, and the use of Boolean delay equation models, the time scale for one cycle of this feedback loop is determined to have a period of about 21 years.

1 Introduction

In recent years much evidence has been presented to suggest the existence of a natural mode of interdecadal climate variability in the subpolar North Atlantic (see Weaver and Hughes 1992 for a review). For example, oxygen isotope records taken from Greenland ice cores (e.g. Hibler and Johnsen 1979) indicate significant variability over the last several centuries with a dominant period of around 20 years. Furthermore, statistical analyses of time series of the various subpolar North Atlantic climatic components (e.g. Deser and Blackmon 1993; Stocker and Mysak 1992) indicate the existence of a mode of variability on the decadal-interdecadal time scale. While it has been known for some time that ocean models can exhibit spontaneous variability on these time scales (e.g. Weaver and Sarachik 1991; Weaver et al. 1991, 1994; Weisse et al. 1994), only recently has such variability been seen in a fully coupled atmosphere-ocean model (Delworth et al. 1993).

Numerous mechanisms have been proposed to explain the variability found in the ocean-only models (e.g. Weaver et al. 1993, 1994; Weisse et al. 1994; Greatbatch and Zhang 1995) as well as simple ice-ocean models (Yang and Neelin 1993; Zhang et al. 1995) and the more complicated atmosphere-ocean-ice models (Delworth et al. 1993). It is still not clear whether or not the variability found in the models can be related to variability found within the true air-sea-ice climate system.

Mysak et al. (1990; see also Mysak and Power 1992) in a seminal study, proposed the existence of a complicated air-sea-ice feedback loop based on a detailed statistical study of the various North Atlantic climatic components. In particular, they suggested that anomalous runoff from the Mackenzie River, perhaps associated with anomalous cyclogenesis over the Greenland-Iceland-Norwegian Seas (GIN) could have been the source for the anomalous freshwater export from the Arctic into the North Atlantic known as the great salinity anomaly (GSA, Dickson et al. 1988). While there is some question as to whether or not enhanced cyclogenesis over the GIN Seas would result in enhanced runoff from the Mackenzie River, it is even less clear whether or not the magnitude of the Mackenzie discharge is large enough to affect the freshwater budget of the North Atlantic. The Mackenzie river has an average annual discharge rate of 7932 $m^3 s^{-1}$ (van der Leeden et al. 1991) which is a factor of four times smaller than the freshwater anomaly associated with the GSA (estimated by Aagaard and Carmack 1989 to be 31 688 $m^3 s^{-1}$ for a period of two years). Thus, anomalies in the seasonal discharge of the Mackenzie River would be of insignificant magnitude compared to

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the magnitude of the GSA without a substantial gain in the system. Furthermore, the Mackenzie River discharge is small compared to the Arctic freshwater transport through the Canadian Archipelago into the North Atlantic (29 153 $m^3 s^{-1}$, Aagaard and Carmack 1989), the Arctic freshwater transport through Fram Strait (114 394 $m^3 s^{-1}$, Aagaard and Carmack 1989), the annual mean St. Lawrence River discharge (14 164 $m^3 s^{-1}$, van der Leeden et al. 1991), and the net runoff into the Atlantic from 45°N to 70°N (121 000 $m^3 s^{-1}$, Baumgartner and Reicher 1975).

One of the purposes of this study is to propose an alternate interdecadal time scale negative feedback loop which is internally self-contained within the subpolar North Atlantic and hence does not rely upon either cyclogenesis in the GIN Sea or anomalous runoff from the Mackenzie River. Our analysis builds upon that of Mysak et al. (1990) as well as statistical analyses which have been conducted since that time.

Lazier (1980), Dickson et al. (1988) Aagaard and Carmack (1989) and Power and Mysak (1992) each discussed the possibility that an anomalous atmospheric high pressure cell over Greenland resulted in an enhanced flux of ice and freshwater through Fram Strait into the Greenland Sea, and hence in the generation of the GSA of 1968–82. Following the reasoning and evidence of Dickson et al. (1988) and Mysak et al. (1990), salinity and ice anomalies are advected out of the Greenland Sea, around the subpolar gyre and eventually enter the Labrador Sea. Fresher surface waters suppressed deep convection, and a reduction in Labrador Sea water (LSW) formation ensued.

In the North Atlantic, the primary source region for North Atlantic deep water (NADW) is the GIN Sea. Winter convection in the Labrador Sea creates an intermediate LSW water mass which overrides the deeper NADW (see for example Lehman and Keigwin 1992). Processes affecting deep water formation or convection in these regions (e.g. the GSA of 1968–82) can have profound effects on both local and global climate. Recent modelling studies (Delworth et al. 1993; Weaver et al. 1994; Weisse et al. 1994) suggest that the source of interdecadal climate variability for the North Atlantic may well lie in the Labrador Sea. Reduced LSW formation could result in a reduction of influx of subtropical waters into the Labrador Sea region (via the North Atlantic and Irminger currents), leading to colder than normal sea surface temperatures (SSTs). Additionally, in the absence of deep convection in the Labrador Sea, surface waters experience enhanced cooling due to sensible and latent heat loss to the atmosphere since mixing with warm deeper waters is limited.

Greatbatch et al. (1991) used the diagnostic model of Mellor et al. (1982) to examine the Gulf Stream transport for the pentads of 1955–59 (warm subpolar SSTs) and 1970–74 (cold subpolar SSTs). Their results suggest the Gulf Stream extension was displaced somewhat southwards during the 1970s and reduced by some 15–30 Sv (1 Sv$=10^6 m^3 s^{-1}$). In a related study, Greatbatch and Xu (1993) used the transport data of Greatbatch et al. (1991) in conjunction with the Levitus (1989a, b, c, 1990) hydrographic data to infer the northward ocean heat transport across 54.5°N and 23.5°N. They suggested the ocean heat transport was some 0.2 PW (1 PW$=10^{15}$ Watts) stronger during the 1950s as compared to the 1970's.

Bjerknes (1964) and Kushnir (1994) argued that while interannual variations in SSTs are governed by local surface wind conditions, interdecadal variations in SSTs are linked to changes in the ocean circulation, and in particular the ocean's thermohaline circulation. Ocean general circulation models further reveal long-time scale oceanic variability associated with changes in the thermohaline circulation and variations in the formation of deep water (see for example Weaver and Hughes 1992 for a review). It is therefore possible that negative salinity anomalies originating in the Greenland Sea [due to positive sea level pressure (SLP) anomalies over Greenland], which are advected into the Labrador Sea via the subpolar gyre, may ultimately affect the SSTs in this subpolar region. If one could show that these subpolar SSTs in turn affect the observed SLPs over Greenland, then a closed negative feedback loop could be proposed for the subpolar North Atlantic.

The outline of the rest of this study is as follows. In the next section we undertake a statistical analysis to show that subpolar SSTs are indeed significantly correlated with the SLPs over Greenland in subsequent years, while in section 3 we provide a physical discussion for the correlation. The relationship between subpolar SSTs and Greenland SLPS, combined with existing analyses available in the literature, allows us to construct an interdecadal negative feedback loop in section 4 analogous to that constructed by Mysak et al. (1990). In section 5 we obtain the time scale of the feedback loop through the use of the lags associated with the correlations between different climatic components, observed horizontal advection time scales, and Boolean delay equation models. A summary is presented in section 6.

2 Correlation of subpolar SST and Greenland SLP

In order to examine the hypothesis that variations in subpolar SSTs may lead to significant variations in SLP over Greenland, a 50 year time series of subpolar SST anomalies (averaged over 40–60°N, 20–50°W) is correlated with a 50 year time series of SLP anomalies (averaged over the Greenland ice cap).

Figure 1a shows the time series of the annual SST anomalies spatially averaged over 40–60°N, 20–50°W extending from 1930 to 1979 (derived from COADS data, Slutz et al. 1985), together with the time series of SLP anomalies averaged over the Greenland ice cap (grid points 65°N, 40°W; 70°N, 35, 40, 45°W; 75°N, 35, 40, 45°W). The SLP anomalies are averaged over the winter months of January and February for the period 1930 to 1979 (data derived from NCAR archived data). For comparison purposes, both series have been nor-

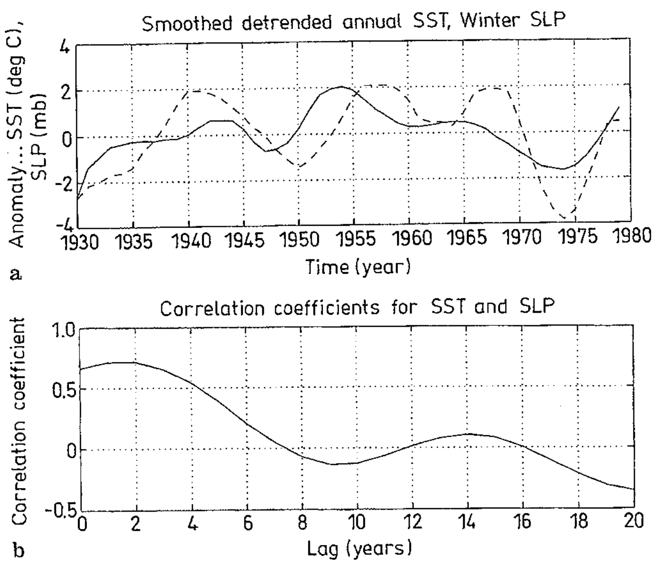

Fig. 1. a The detrended 9 year running means of the SST and SLP anomalies are plotted together for visual comparison. Note that the magnitudes of the SST anomalies have been multiplied by 5 for better visual correlation. (*Solid line,* SST; *Dashed line,* SLP). **b** The correlation coefficients between the two time series at various lags extending from 0 to 20 years, where SST leads SLP. The SST anomaly time series is derived from COADS data (Slutz et al. 1985) while the SLP anomaly time series is derived from archived NCAR data. Both the SST and SLP data were obtained from Dr. Shiling Peng

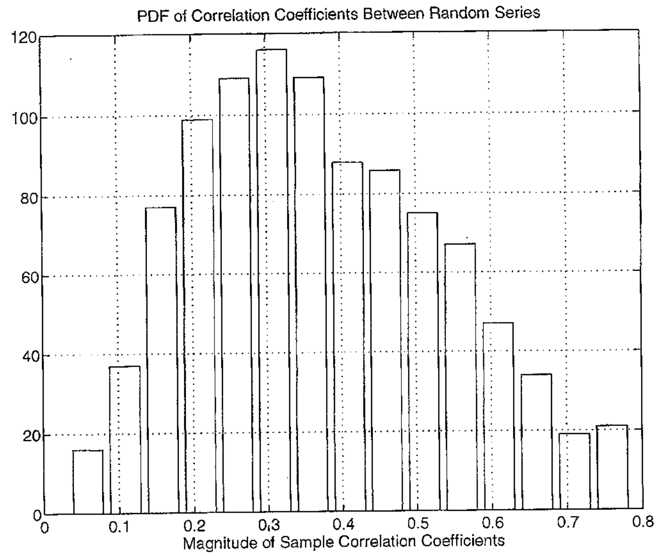

Fig. 2. Histogram of the maximum 1000 coefficient magnitudes found within the range of lags extending from $\tau = -2$ to $\tau = +2$

malized, detrended, and smoothed with a 9 year weighted running mean in order to remove variations occurring on less than decadal time scales.

The maximum correlation between the two time series occurs at a lag of 2 years (SST leading SLP), with a correlation coefficient of 0.72. A lag of 1 year is to be expected as "winter" in the SLP anomaly time series has been defined to consist of the months of January and February. Some dependence of SLP upon the SST anomalies of the previous December (in the previous year) is to be expected. To show this, the winter time series of SST (not shown) was correlated with the winter time series of SLP. In this case, the lag disappears, and the maximum correlation coefficient is 0.71.

Figure 1b shows the correlation coefficients between the two time series plotted versus the lag (in years). A general peak occurs over the lags of 1 and 2 years with the coefficient at 2 years (0.716) being slightly higher than that at 1 year (0.715). This smearing of the peak over adjacent lags is to be expected since the SST anomaly time series is averaged over such a large area, the different grid points of which exhibit peaks at slightly different times. Time series of annual SST anomalies at Ocean Weather Ship (OWS) Bravo, situated in the Labrador Sea (Lazier 1980; Michaud and Lin 1991), and at OWS Charlie, situated near 35.5°N, 54.5°W (Levitus et al. 1994), show that the peaks in SST at OWS Bravo lead the peaks in SST at OWS Charlie by approximately two years. Averaging over the large area 40–60°N, 20–50°W will average such lags as well.

In order to test the significance of a correlation coefficient $r = 0.72$ occurring at a lag of 2 years, extra care was taken in order to account for the reduction in the number of degrees of freedom due to the inherent autocorrelations of the time series involved, and in light of the smoothing applied to each time series in order to remove variations occurring on time scales of less than 10 years. Two sets of 1000 random time series (of length $n = 50$) were generated using a first order Markov process to give each time series an autoregressive nature, one set having the characteristics of the SST time series, and the second set having the characteristics of the SLP time series (before smoothing and detrending). Once the random Markov series were generated, they were then smoothed and detrended (processed just as the original time series were processed) and then correlated. The correlation coefficients between these random time series were then calculated, and the probability distribution function (pdf) of these 1000 random correlation coefficients was plotted. Since the significance of a correlation coefficient of $r = 0.72$ at a lag of $\tau = 2$ years is being tested, the absolute value of the maximum correlation coefficient in each case which was found to lie within the range of lags from -2 years to $+2$ years is plotted in the histogram of Fig. 2. From this figure, it is evident that the probability of a correlation coefficient of magnitude 0.7 or greater occurring within the lags $\tau \leq |2|$ is equal to $\frac{40}{1000} = 4\%$. Thus a correlation coefficient of $r = 0.72$ at lag $\tau = 2$ is 96% significant. This suggests that 52% of the hindcast variance can be explained through a simple linear regression.

The correlation between annual SST and winter SLP anomalies was repeated with the end-points of the time series removed. Due to the fact that a 9-point weighted running mean was used to smooth the series, the end points are biased to one side. Leaving off the first and last 2 points of the time series before correlat-

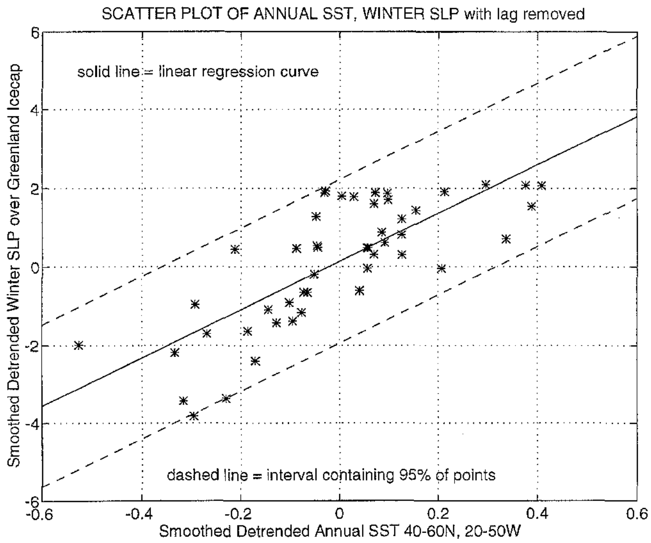

Fig. 3. Scatter plot of Greenland SLP anomalies versus subpolar North Atlantic SST anomalies, once the 2-year lag between the two has been removed. The *dashed lines* are the linear regression line $\pm 2 \times$ (the standard error of estimate). The interval between these lines should, if the number of points, n, is large enough, contain approximately 95% of the data points

ing them gives a correlation coefficient of $r=0.74$ at lag $\tau=2$ years (SST leading SLP). From the preceding method of determining significance, this correlation coefficient (for $n=46$ points) is still found to be 95% significant. Leaving off the first and last 4 points of the time series before correlating gives a correlation coefficient of $r=0.67$ at lag $\tau=2$ years. However, in this case we are no longer left with a time series of sufficient length in order to achieve results significant at the 95% confidence level.

Some dependence of the SLP over Greenland upon the SSTs of the subpolar North Atlantic may therefore be expected, since the SST anomalies lead the SLP anomalies. In order to investigate this possibility further, the lag between the two time series was removed, and the SLP anomalies were plotted against the SST anomalies in a scatter plot (see Fig. 3). A linear regression curve fitted to the scatter points gives

$$SLP(mb) = 6.1470 \times (SST(^\circ \mathrm{C})) + 0.1355. \quad (1)$$

This linear regression gives a larger gain in SLPs for a given change in SSTs than was suggested by the modelling and observational study of Palmer and Sun (1985), however it explains the data it was modelled from well. For example, SSTs in the subpolar gyre (focused in the Labrador Sea) dropped approximately 1° C from the 1950s to the early 1972s (Peng and Mysak 1993). Figure 1 indicates that over the period 1955–59 to 1970–74, the SLPs over Greenland fell by approximately 6 mb. The results of Palmer and Sun (1985) suggest only a 0.5 mb response in SLPs over Greenland for a given 1° C in SST, with these results being less than 95% significant. Notice however that Palmer and Sun (1985) used SLP data over a wider region including the ocean and so a detailed comparison with our results is not possible.

3 Physical discussion of the correlation between subpolar SST anomalies and Greenland SLP anomalies

It was concluded in the previous section that subpolar SSTs have a statistically significant impact upon SLPs over Greenland, although the actual magnitude of this impact requires further study. In this section, the question of *how* subpolar SSTs are able to affect SLPs over a remote region such as Greenland is addressed.

Subpolar SSTs may affect SLPs over Greenland in two different ways: The first was suggested by Dickson (1994) where he proposed that a strengthening of the statistical atmospheric high (or ridge of high pressure) which dominates over the Greenland ice cap may simply be viewed as a weakening of the statistical Iceland Low, and vice versa. The climatological Iceland Low is found to vary significantly with variations in the climatological storm tracks (Ueno 1993). A weak North Atlantic Oscillation (NAO) pattern with a weak Iceland Low and a weak atmospheric jet, is significantly correlated with storm tracks whose main routes are oriented eastward along 45–50° N reaching the west of Europe. A strong NAO pattern with a strong Iceland Low and a strong atmospheric jet is correlated with cyclones of strong intensity which follow tracks northeastward, passing along the south coast of Greenland to reach the Norwegian Sea (Ueno 1993). The atmospheric jet and associated storm tracks, in turn, are affected by the north-south temperature gradient across the north wall of the Gulf Stream (Dickson and Lee 1969; Palmer and Sun 1985). A negative (positive) SST anomaly in subpolar waters leads to enhanced (weakened) oceanic and atmospheric baroclinic zones in the vicinity of the Gulf Stream, and hence greater (less) than normal amounts of available potential energy for developing baroclinic systems. Atmospheric lows or storms will therefore be more intense (weaker) and generate greater (less) than normal amounts of cyclonic vorticity. This excess (lack of) vorticity facilitates (opposes) the northward deviation of storm tracks away from the climatological mean tracks. Also, as noted by Kushnir (1994), Peng and Mysak (1993), Deser and Blackmon (1993), and Palmer and Sun (1985), negative SST anomalies in the Labrador Sea region are associated with a general anticyclonic circulation over the subpolar gyre area, while positive SST anomalies are associated with a general cyclonic atmospheric circulation over the subpolar gyre region (downstream from the main focus of the SST anomaly). The atmospheric jet and associated storm tracks will therefore be affected by these circulation patterns, which favour the northward movement of storms during negative SST anomaly years, and oppose the northward movement of storms during the positive SST anomaly years. Palmer and Sun (1985), in their modelling and observational study concerning the atmospheric response to subpolar SST anomalies, found some indication of enhanced eddy variations over Greenland (to the northeast of the area 40–50° N, 40–60° W) although these results were less than 95% significant. These authors also noted that the magnitude of the atmospheric height difference downstream of the SST anomaly appears to depend approximately linearly upon the magnitude of the SST anomaly (similar to the response found in the correlation in the previous section).

The second way in which subpolar SST anomalies are able to affect SLPs over Greenland is through thermal responses. Warmer than normal SSTs surrounding Greenland enhance the density difference between the air above the cold land mass and the air overlying the ocean. This, in turn, enhances the outflow of cold dense air from land to ocean, and generates anomalous subsidence and negative vorticity over the ice cap, both of which serve to strengthen the Greenland High. Conversely, colder than normal SSTs surrounding Greenland weaken the density difference in the atmospheric between the air overlying the land and the air overlying the water, resulting in conditions which serve to weaken the Greenland High.

4 Proposed decadal climate loop for the subpolar North Atlantic

As a result of the analysis discussed in the preceding sections we now propose the interdecadal negative feedback loop for the subpolar North Atlantic shown in Fig. 4. This feedback loop is analogous to the one proposed by Mysak et al. (1990) except that we do not rely upon GIN Sea cyclogenesis and Mackenzie River runoff to explain ice export through Fram Strait. Notice that we have included the complete feedback loop in Fig. 4 whereas Mysak et al. (1990) and Mysak and Power (1992) show only half of their climate cycle.

To summarize: a positive SST anomaly in the subpolar North Atlantic (associated with deep convection in the Labrador Sea and a strong circulation of North Atlantic Current waters into the Labrador Sea region via the Irminger Current) is positively correlated with higher than normal SLPs over Greenland. The reason for this is that positive SST anomalies divert storm

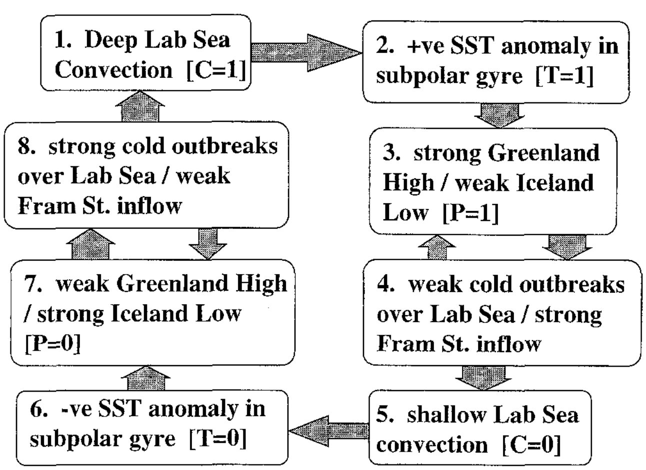

Fig. 4. Proposed interdecadal climate cycle for the subpolar North Atlantic

tracks south of their normal position, resulting in higher SLPs in the north (stronger climatological high over Greenland and weaker climatological Iceland Low). This leads to a greater than normal flow of ice and fresh water through Fram Strait into the Greenland Sea (Dickson et al. 1988; Aagaard and Carmack 1989; Power and Mysak 1992). A weak Iceland Low is also associated with weak atmospheric circulation over the subpolar North Atlantic. Both of these factors, but primarily that of a developing negative salinity anomaly (SA) in the Greenland Sea which advects around the subpolar gyre into the Labrador Sea, serve to reduce the maximum convective limit in the Labrador Sea in winter. The circulation of warm, saline North Atlantic Current waters into the Labrador Sea is reduced, and cooler than normal SSTs in the subpolar gyre result.

These cooler than normal SSTs serve to divert storms northward of their climatological track, lowering SLPs in the north and hence weakening the Greenland High and enhancing the Iceland Low. The inflow of ice and freshwater through Fram Strait is therefore reduced, and the bulk of the SA in the Labrador Sea now advects out of the region. An enhanced Iceland Low also serves to produce enhanced cold winter atmospheric outbreaks over the Labrador Sea. Both the advection of the negative SA out of the central Labrador Sea region and the increase in winter heat loss due to the increased atmospheric cold outbreaks over the area serve to re-initiate deep convection in the Labrador Sea. The branch of the North Atlantic Current into the Labrador Sea is therefore enhanced, and warmer than normal SSTs in the subpolar region result once again.

It should be noted that there are two feedbacks or sub-loops within the main cycle. The first occurs between boxes 3 and 4, and the second occurs between boxes 7 and 8. Briefly, in the first sub-loop, higher than normal pressures over Greenland drive increased amounts of ice and fresh water through Fram Strait into the Greenland Sea. This ice and fresh water acts to reduce temperatures in this northern region, and thermally enhance the Greenland High, resulting in an even greater transport through Fram Strait of ice and fresh water. Thus, the Greenland High builds in strength as events cycle through this sub-loop. The building negative SA in the region begins to advect around the subpolar gyre and approximately four years after the beginning of the higher SLPs in the north, convection in the Labrador Sea begins to shallow (and subpolar SSTs begin to cool). The climate breaks out of the sub-loop once the bulk of the Greenland Sea SA has been advected into the Labrador Sea and LSW formation has ceased. This follows since SSTs have cooled such that storm tracks begin to divert northward of their climatological track and flow through Fram Strait begins to decrease.

The second feedback is described as follows: once storm tracks are diverted north of their normal position by cooling SSTs in the subpolar gyre, and the bulk of the SA has been advected out of the central Labrador Sea, deep convection once again resumes in the

Labrador Sea. This increases the meridional overturning of the thermohaline circulation, and increased heat begins to be transported northward once again by the Gulf Stream. However, the more northerly storm tracks deepen the Iceland Low, and produce stronger cold atmospheric winter outbreaks over the Labrador Sea, enhancing and continuing the cooling of SSTs in the region (even though the recirculation branch of the North Atlantic Current into the Labrador Sea is beginning to increase the heat transport into the region). This serves to enhance the temperature contrast across the north wall of the Gulf Stream and further strengthen the atmospheric baroclinic zone, producing more frequent storms of greater depth in central pressure. Roebber (1984) found more explosively deepening cyclones off of Cape Hatteras and northeastward in 1979–1982 than in 1976–1979. Also, Thompson and Hazen (1983) show that the Iceland Low of winter 1972 was weaker than in winter 1974. Both of the above references show an increasing trend in storm frequency and depth once the GSA passed out of the Labrador Sea and deep convection was renewed. This is only possible if an increased baroclinic zone existed due to an increasing temperature gradient across the north wall of the Gulf Stream. These storms, due to greater than normal amounts of vorticity generation, travel along tracks even further north of the climatological track, further enhancing the Iceland Low. The cold outbreaks over the Labrador Sea are thus increased and hence so is the cooling of surface waters. The system breaks out of this sub-loop once the oceanic transport of heat into the Labrador Sea becomes sufficient (and has had sufficient time) to once again raise SSTs above their climatological norm, and storm tracks have once again been diverted south of their normal tracks.

5 Time scale of proposed subpolar North Atlantic climate loop

The time scale of the negative feedback loop is determined here, through the use of a nonlinear Boolean delay equation (BDE) model (as in Darby and Mysak 1993). For simplicity, the following 3 Boolean variables are substituted for the separate climate components discussed above as part of the negative feedback loop: (1) $C \equiv$ Labrador Sea convection (where $C=1$ represents deep convection; $C=0$ represents shallow convection); (2) $T \equiv$ Subpolar SST anomaly (where $T=1$ represents a positive anomaly and $T=0$ represents a negative anomaly); and (3) $P \equiv$ Greenland SLP anomaly [where $P=1$ represents higher than normal atmospheric sea level pressures in the north (i.e. strong Greenland High, weak Iceland Low) as well as the associated states of strong Fram Strait inflow of ice and fresh water into the Greenland Sea, and weak cold outbreaks over the Labrador Sea; $P=0$ represents lower than normal pressures in the north (i.e. weak Greenland High, strong Iceland Low) as well as the associated states of weak Fram Strait inflow, and strong cold outbreaks over the Labrador Sea].

Maximum (minimum) subpolar SSTs appear to follow maximum (minimum) winter convective depths in the Labrador Sea by approximately 2 to 3 years. For example, in Fig. 1 it is seen that a minimum in SSTs occurred in 1974 (in the 9-year running mean). Minimum convection due to the passage of the GSA through the Labrador Sea occurred in 1971 (Lazier 1980). If this is taken for the norm, then

$$T(t)=C(t-3). \tag{2}$$

From section 2 the correlation between subpolar SSTs (T) and Greenland SLPs (P) suggests that SSTs lead the SLPs by 1 year:

$$P(t)=T(t-1). \tag{3}$$

From Dickson et al. (1988), the advection time for a salinity anomaly around the subpolar gyre from the Greenland Sea to the Labrador Sea is approximately 4 years, so that

$$\overline{C(t)}=P(t-4), \tag{4}$$

where the overbar signifies the logical operator "not". That is, convection in the Labrador Sea depends upon the atmospheric state over Greenland 4 years previously. If the sea level pressure was anomalously high ($P=1$) 4 years ago, then a negative salinity anomaly will have been generated in the Greenland Sea and had time to advect around the subpolar gyre into the Labrador Sea, weakening convection ($C=0$).

Equation 4 can not be used in isolation since we must also add in the fact (through the use of the logical operator "and") that convection in the Labrador Sea is also dependent upon the atmospheric state at the current time:

$$C(t)=P(t). \tag{5}$$

That is, convection in the Labrador Sea depends upon the existence of outbreaks of cold air from the northern North American continent over the relatively warm Labrador Sea. When SLPs over Greenland are anomalously low, the Iceland Low is anomalously deep (as was discussed in section 3). This means that storms or lows developing along the main atmospheric baroclinic zone at mid-latitudes over the North Atlantic are more intense and frequent, following storm tracks north of the climatological mean. Thus, when $P=0$, cold outbreaks over the Labrador Sea in the wake of these storms are also more intense and frequent than normal. These cold outbreaks enhance the sensible and latent heat loss from the ocean surface to the atmosphere, and therefore enhance convection. Conversely, when $P=1$, cold outbreaks over the Labrador Sea are reduced below normal in intensity and frequency, resulting in an atmospheric pattern which is less favourable to the cooling of the ocean surface and convection.

Deep convection in the Labrador Sea is then the result of no current negative salinity anomalies affecting the region at the surface ($P(t-4)=0$), *and* strong cold

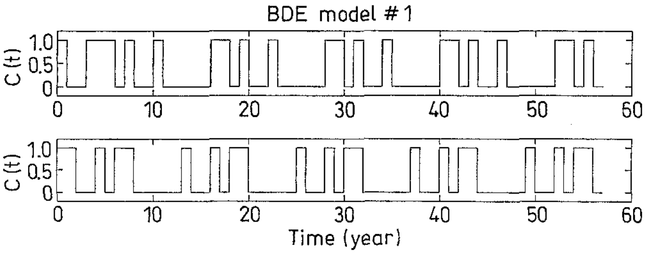

Fig. 5. Results from the first nonlinear BDE model, for two randomly selected initial conditions, showing a 12-year cycle in convection ($C(t)$) in the Labrador Sea

atmospheric outbreaks ($P(t)=0$). Shallow convection is the result of a negative salinity anomaly passing through the region ($P(t-4)=1$), *or* weak cold atmospheric outbreaks ($P(t)=1$). The model equations are thus:

$$C(t)=\overline{P}(t-4) \wedge \overline{P}(t) \tag{6}$$

$$\overline{C}(t)=P(t-4) \vee P(t) \tag{7}$$

(where $\wedge$ denotes the logical "and", and where $\vee$ denotes the logical "or"). The these equations are identical, and either may be used as the basis for our BDE model.

From Eqs. 2 and 3, Eqs. 6 and 7 become

$$C(t)=\overline{C}(t-8) \wedge \overline{C}(t-4) \tag{8}$$

$$\overline{C}(t)=C(t-8) \vee C(t-4) \tag{9}$$

As noted before, these equations are identical and either may be used as the BDE model. Two examples (using different random initial conditions) of the results are shown in Fig. 5. While using different random initial conditions gives different structure to the cycles, all resulting cycles have 12-year periods. This must be so, since $\overline{C}(t-8) \wedge \overline{C}(t-4)$ was prescribed which leads to the summation of the time lags $\tau=8$ and $\tau=4$ to give a 12-year time lag.

It has been noted that in the shallowing of convection in the Labrador Sea, negative SAs are far more effective than are weak cold outbreaks. Equation 9 could therefore be shortened to

$$\overline{C}(t)=C(t-8). \tag{10}$$

Since this equation is no longer identical to Eq. 8, the following method is adopted in order to apply the equations as one model:

If $C(t-1)=1$, then $C(t)=1$, *unless* $C(t-8)=1$.
If $C(t-1)=0$, then $C(t)=0$, *unless* $C(t-8)=0$ *and* $C(t-4)=0$.

From this, 12 year cycles are still obtained (not shown), however with different structures from those displayed in the previous figure.

These 12 year cycle periods obtained in the BDE model fall short of the expected 20–22 year climate variations observed in the North Atlantic (e.g. Hibler and Johnsen 1979). Thus, a revised version of the BDE model, as an extension of the above model, is formulated using the observation that a trend of increasing pressures in the north is required in order to generate a negative salinity anomaly of significant magnitude to shallow convection in the Labrador Sea (see Dickson et al. 1975; Häkkinen 1993). The time series of SLP anomalies over Greenland during the 1960s indicates that over the 8-year period extending from 1962 to 1969, the SLPs were normal or above normal in magnitude for 5 of those 8 years. This period exhibits three individual peaks (which have been smoothed out in Fig. 1), with each peak greater in magnitude than the last such that in the 9-year weighted mean, a rising trend in SLPs over Greenland is noted. Thus, the following may be prescribed as a necessary condition in order to turn convection "off". If $C(t)$ is "on", then it will stay on ($C=1$), unless 5 of the preceding 8 years (with an additional 4 years added on to each to account for the time necessary for a generated SA to reach the Labrador Sea) experienced strong ($P=1$) pressures over Greenland. That is, if $C(t-1)=1$, then $C(t)=1$, *unless* at least 5 of

$$\left\{\begin{matrix} P(t-4) \\ P(t-5) \\ \vdots \\ P(t-11) \end{matrix}\right\} = 1$$

or, using the fact that $T(t)=C(t-3)$, and $P(t)=T(t-1)$,
if $C(t-1)=1$, then $C(t)=1$, *unless* at least 5 of

$$\left\{\begin{matrix} C(t-8) \\ C(t-9) \\ \vdots \\ C(t-15) \end{matrix}\right\} = 1.$$

The second condition needed for the model concerns the renewal of deep convection in the wake of an SA, or $C=0$. Here, it is possible to use the same equation used in the first model. That is,

If $C(t-1)=0$, then $C(t)=0$, *unless* $C(t-8)=0$ *and* $C(t-4)=0$.

One should note that since different conditions apply depending upon the state of convection in the Labrador Sea, it is necessary to separate these equations or conditions as in the second version of the first BDE model. First, the current state (at time $t-1$) is tested to see whether convection is on or off. If it is on, then as in the first condition above, it will not turn off unless 5 years of higher than normal pressures are experienced in the 8 years extending from $t-4$ to $t-11$. If convection is off, then the conditions required to turn it back on are: (1) no SAs during the current year, and (2) strong cold outbreaks or lower than normal pressures in the north.

Figure 6 shows the results from two integrations of this model using two randomly selected different initial conditions. Notice that although two different sets of random initial conditions were used, the same resultant structures in the cycles are realized, both with a period of 21 years. The structures of these cycles appear much

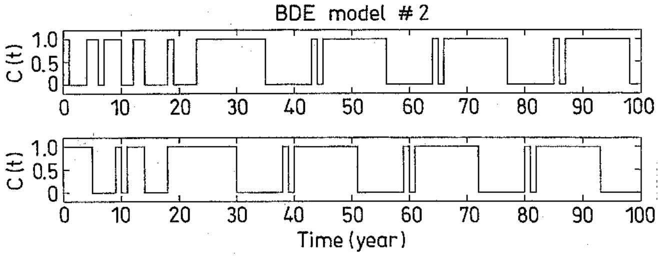

Fig. 6. Results from the first extended nonlinear BDE model, for two randomly selected initial conditions, showing a 21-year cycle in convection ($C(t)$)

more reasonable than do those of the first BDE model, and hence we conclude that this model is more appropriate. (Note: if it is required that all 8 of the 8 years experience $P=1$, then 23-year cycles are obtained – not shown here).

One of the main conclusions drawn from this investigation is that the build-up period of the Greenland High is of key importance in determining the time scale for the entire feedback loop, and in lengthening subpolar north Atlantic climate variations from decadal to interdecadal timescales.

6 Summary and conclusions

The main contribution of this research is the finding that subpolar North Atlantic SST anomalies do have an impact upon atmospheric SLP anomalies over Greenland (as was independently suggested by Dickson 1994). Extra care was given to account for inherent autocorrelations within time series of the different climate components and the significance tests of the correlation coefficients were adjusted for reduced numbers of degrees of freedom.

As a result of this correlation and other statistical relationships discussed in the literature, the following interdecadal climate cycle was proposed. Positive SST anomalies concentrated in the vicinity of the Labrador Sea reduce the north-south temperature gradient across the north wall of the Gulf Stream, and hence the north-south temperature gradient in the air overlying this ocean current. This weakens the atmospheric baroclinic zone and hence the accompanying jet stream, such that a smaller than normal amount of available potential energy (APE) is available for developing baroclinic eddies or storms. These storms or lows have greater than normal central pressures and generate less than normal local vorticity. This lack of vorticity constrains the storms to follow paths south of their climatological tracks, and hence causes SLPs in the north to rise to magnitudes above normal, enhancing the climatological Greenland High and weakening the climatological Iceland Low. The anomalous anticyclonic wind field generated by the stronger than normal Greenland High enhances the transport of freshwater and ice through Fram Strait from the Arctic Ocean into the Greenland Sea. This creates a negative salinity anomaly at the sea surface of the Greenland Sea, which is then advected around the subpolar gyre into the Labrador Sea. The lower than normal sea surface salinities (SSSs) serve to reduce LSW formation and hence reduce the amount of North Atlantic Current water entering the subpolar gyre and Labrador Sea via the Irminger Current. This further reduces the SSTs and SSSs, in the region.

Negative SST anomalies concentrated in the Labrador Sea region of the subpolar North Atlantic ocean gyre serve to enhance the north-south temperature gradient across the north wall of the Gulf Stream, and thus also enhance the accompanying atmospheric baroclinic zone and jet. Developing atmospheric baroclinic eddies or storms therefore have access to greater than normal amounts of APE, and attain central pressures which are deeper than normal. These storms generate larger than normal amounts of local vorticity which allows the lows to travel northeastward along tracks north of the normal position, enhancing the climatological Iceland Low and reducing the Greenland High. This results in an anomalous wind stress pattern unfavourable for the export of ice and freshwater from the Arctic Ocean through Fram Strait into the Greenland Sea, allowing sea surface salinities in the Greenland Sea to return to normal or above (due to the import of salt into the region by the North Atlantic Current). This anomalous atmospheric circulation pattern also favours an increase in strong outbreaks of cold continental air from the Canadian Arctic Archipelago over the relatively warm Labrador Sea, further supporting the renewal of deep water formation or convection. This increases the amount of North Atlantic Current water entering the subpolar gyre via the Irminger Current, and once again serves to raise SSTs above normal in the vicinity of the Labrador Sea.

Through the time lags associated with the correlations between the various climate components, known horizontal advection time scales in the subpolar gyre, and the use of a Boolean delay equation model, the time scale for the proposed interdecadal climate cycle was determined to have a period of approximately 21 years. The time during which the Greenland High remains above normal in strength (or the time during which it builds up in magnitude) was found to play a key role in determining the time scale of the feedback loop. This time scale agrees well with the 20 year climate variability found in ice cores (Hibler and Johnsen 1979), and in EOF analyses of different fields of the various North Atlantic climatic components (e.g. Deser and Blackmon 1993).

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