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IMPACT OF LABRADOR SEA-ICE EXTENT ON THE NORTH ATLANTIC OSCILLATION

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ABSTRACT

The wintertime atmospheric response to imposed sea-surface temperature and sea-ice extent changes in the Labrador Sea has been investigated by means of ensemble simulations with an atmospheric general circulation model. Low temperatures and heavy ice conditions in the Labrador Sea produce a statistically significant (at 95% confidence) negative North Atlantic oscillation–Arctic oscillation (NAO–AO) response. Conversely, reduced sea-ice extent in the Labrador Sea produces a positive NAO–AO response. The two simulations with opposite sea-ice conditions in the Labrador Sea exhibit a maximum mean wintertime difference of 4–5 hPa in sea-level pressure corresponding to a substantial and statistically significant change in the NAO–AO index of 0.7 standard deviations. The large-scale response to a local perturbation of sea-ice conditions is associated with marked changes in the transient eddies (synoptic storms). Changes in the sea-ice cover cause changes in low-level baroclinicity that perturb the travelling baroclinic disturbances, which then bring the signal downstream to manifest a non-local Atlantic-wide response. The atmospheric response suggests that the sea ice in the Labrador Sea is able to provide an important negative feedback on long-term NAO–AO variations. Copyright © 2004 Royal Meteorological Society.

KEY WORDS: Atlantic variability; North Atlantic oscillation–Arctic oscillation; storms; Labrador Sea; sea ice

1. INTRODUCTION

Sea ice is an important component of the climate system that affects the atmosphere through surface albedo, exchange of heat, moisture and momentum between the atmosphere and the ocean. Sea ice also influences the upper ocean stratification and is important for deep-water formation. Sea ice is, in turn, strongly influenced by oceanic and atmospheric circulations. Deser *et al.* (2000) described the Arctic sea-ice extent variability during the past 50 years and noted that the largest variability occurs in winter over the Atlantic sector. The leading mode of wintertime sea-ice variability was found to be a see-saw pattern with the centres of action in the Labrador Sea and Greenland–Iceland–Norwegian (GIN) Seas. In addition, Deser *et al.* (2000) showed that the leading mode of variability explained 35% of the total variance and had a significant correlation with the wintertime North Atlantic oscillation (NAO) index of 0.69. The high (low) phase of the NAO is associated with an extensive (reduced) sea-ice cover in the Labrador Sea and reduced (increased) sea-ice cover in the GIN Seas. Closer investigations of lead/lag relationships indicate that Arctic sea-ice (both in the Labrador and GIN Seas) is responding to atmospheric forcing on monthly to interannual time scales (Prisenberg *et al.*, 1997; Deser *et al.*, 2000). On longer time scales, however, these relationships are more debatable, and in this case it is relevant to address the atmospheric response to sea-ice anomalies, since such feedbacks are likely to shape longer term secular changes in climate (Deser *et al.*, 2000).

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Far more published studies have used atmospheric general circulation models (AGCMs) to investigate the impact of sea-surface temperature (SST) rather than sea-ice on the atmosphere (Kushnir *et al*., 2002). Lopez *et al*. (2000) was one of the few studies that investigated the atmospheric response to changes in North Atlantic sea-ice cover. However, Lopez *et al*. (2000) simulated the atmospheric response to a complex, nonlocal combined SST and sea-ice anomaly in the North Atlantic. A series of four perturbation experiments were carried out with different sign combinations of the pattern poles in the SST–sea-ice anomaly. The analysis of the simulated atmospheric responses indicated that Labrador Sea surface conditions were the main controlling factor in producing the atmospheric response of the four patterns. However, this was not fully tested by performing a separate experiment with a single localized anomaly in the Labrador Sea. This is a necessary, but probably not sufficient, experiment to test such a hypothesis, since the atmospheric response may be nonlinearly dependent on the shape, amplitude and position of a high/mid-latitude SST anomaly (Kushnir *et al*., 2002). Furthermore, the SST anomaly used in Lopez *et al*. (2000) was almost as large as basin scale, and only relatively small parts of it involved sea-ice-cover changes. Lopez *et al*. (2000) did not quantify the relative contribution to the atmospheric response from the sea-ice and SST parts of the anomaly. It may be assumed that the sea-ice part contributed significantly to the response, since changes in sea-ice cover have a larger impact on the surface energy fluxes than SST changes in already-open sea areas. Furthermore, earlier studies have generally demonstrated only a weak response to North Atlantic SST anomalies (Kushnir *et al*., 2002; Paeth *et al*., 2003).

In this study, we investigate the atmospheric response to a sea-ice-cover anomaly in the Labrador Sea. The Labrador Sea is a key part of the climate system (The Lab Sea Group, 1998) due to the large air sea fluxes taking place in this region and the associated deep vertical mixing of water masses. A recent model study by Bentsen *et al*. (in press) identified a close link between the deep vertical mixing in the Labrador Sea and the variability of the Atlantic meridional overturning circulation (AMOC). They also noted that the ocean convection in the GIN Seas does not show the same close connection to the AMOC variability on the time scales examined. These relations further underline the importance of quantifying the atmospheric effect of surface conditions in the Labrador Sea.

The following section describes the experimental design. In Section 3, we present model results for both mean flow and transients. In general, the mid-latitude transients play a large role for the general circulation (Holopainen, 1990). Also, large differences between the full AGCM response and the direct linear response to mid-latitude SST anomalies are often due to the transient eddy response in the full system (Kushnir *et al*., 2002). Therefore, we provide a more detailed description of the transient eddy behaviour in our experiments in Section 3. However, we have not attempted to quantify the transient eddies' impact on the large-scale flow. Section 4 contains the conclusions.

2. EXPERIMENTAL DESIGN

The model employed in this study is the ARPEGE/IFS model, used operationally by Météo-France and documented in Déqué *et al*. (1994) and Doblas-Reyes *et al*. (1998). The horizontal truncation used here is a linear T63 truncation ($T63_L$), with a lat/lon grid spacing of about 2.8° that has been reduced in the longitude direction near the poles. There are 31 levels in the vertical, with 20 in the troposphere and 10 in the stratosphere.

The sea-ice boundary is defined in ARPEGE by the $-1.9\,^\circ$C SST isotherm, which means that sea ice can be perturbed by perturbing the SSTs in the model. To obtain a local SST–sea-ice anomaly, a 16 year long time series of monthly mean SSTs in the Labrador Sea (60–50 °W, 55–60 °N) was extracted from the AMIP dataset (Gates, 1992). From this time series, an index was constructed of area-averaged SST anomalies (monthly SST minus the long-term mean) for the calendar months November through to March in the period 1979–95. A local positive SST anomaly was subsequently made by regressing global SST anomalies onto the index. The amplitude of the anomaly decreases with distance out from its centre (index area). To remove weak, remote anomalies, the anomaly has been set to zero when the correlation coefficient (between the index and the global anomalies) is less than 0.34. The remaining anomaly was multiplied by ± 3, corresponding to ± 3 standard deviations of the SST index and then superposed on the January, February and March (JFM) climatological SST dataset (Reynolds and Smith, 1994) (Figure 1). The SST index spans five standard deviations during these 16 winters analysed (80 realizations), and one standard deviation corresponds to an SST anomaly of about 1 °C in the area of highest variability.

The model experiments were designed as follows. First, one 14 year control simulation (CTRL) was performed with mean climatological, seasonally varying SSTs. The mean annual cycle of the SSTs is constituted by monthly means that are interpolated linearly in time and updated daily. Two extra annual SST cycles (LABMAX and LABMIN) have been made in addition to the climatological dataset by Reynolds and Smith (1994). LABMAX was made by subtracting the SST anomaly described above from the climatological dataset. In this way we obtained a localized positive sea-ice-extent anomaly in the Labrador Sea with climatological SSTs and sea ice elsewhere. LABMIN was constructed similarly by adding the localized SST anomaly to the climatological SSTs. From CTRL, 14 initial states from 14 different November months were extracted. Two winter (November–March) experiments were carried out, each consisting of 14 runs. The two winter experiments consist of one ensemble with maximal (LABMAX) and one with minimal (LABMIN) Labrador sea-ice extent (Figure 1). For each ensemble member the integration period before January is considered as spin-up time. This ensemble procedure is the same as used by Lopez *et al.* (2000). However, the location, scale and strength of the SST–sea-ice anomaly are quite different in this study. The shape of the SST–sea-ice anomaly used in Lopez *et al.* (2000) was selected by searching the GISST 2.2 dataset (Rayner *et al.*, 1996) for particularly cold and warm winters in the North Atlantic. The SST pattern selected was a nonlocal SST tripole, whereas our anomaly is localized over the Labrador Sea. Compared with present multi-annual SST anomalies, the anomaly in Lopez *et al.* (2000) was amplified by a factor of 5–6 in order to get conditions more similar to those believed to have taken place in the Little Ice Age. That resulted in maximum SST amplitudes of around 5 °C, whereas the maximum amplitude in our study is around 3 °C.

Since the large sea-ice variability in the Labrador Sea seems to be a part of an east–west dipole pattern on the monthly to interannual time scales (Deser *et al.*, 2000), isolation of the surface conditions in this area could be criticized as being unrealistic. However, the pattern in Deser *et al.* (2000) explains 35% of the sea-ice variance, so a considerable amount of variability might be explainable by this localized feature. In addition, long-term trends in sea-ice cover, not apparent in the short observational records, can exist independently of the observed short-term dipole pattern. The long time scales may be of particular relevance here since the atmospheric feedback is likely to take place on these time scales (Deser *et al.*, 2000).

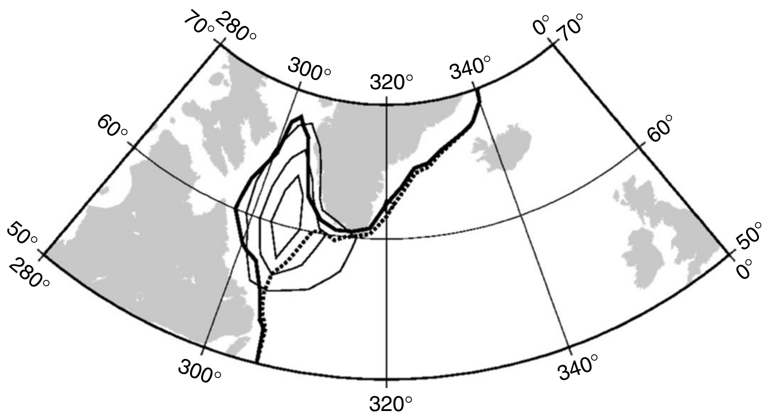


Figure 1. Thin solid contours show Labrador Sea SST anomaly (contour interval 1 K). Thick dotted contour shows the perturbed sea-ice border for January when the SST anomaly has been subtracted from climatology (LABMAX). Thick solid contour shows perturbed sea-ice boundary for January when the SST anomaly has been added to climatology (LABMIN)

3. RESULTS

3.1. Mean response

The winter mean (JFM) difference between LABMAX and LABMIN in geopotential height at 1000 hPa (Φ_{1000}) is shown in Figure 2(a). The corresponding differences between each of the perturbed experiments and CTRL are not shown. However, these responses have a similar pattern as the field in Figure 2(a), but with weaker amplitude (and opposite sign in the LABMIN − CTRL case). It can be seen in Figure 2(b) that the response is nearly equivalent barotropically, since the pattern of the response in Φ_{500} is similar to the Φ_{1000} response. The patterns show a striking resemblance to the NAO–Arctic oscillation (AO) pattern (Hurrell, 1995; Thompson and Wallace, 1998; Ambaum *et al.*, 2001). A two-sided Student t-test shows that substantial areas of the response fields are statistically significant (at the 5% level). To summarize, our results show that a local sea-ice anomaly in the Labrador Sea has created a statistically significant NAO–AO response.

3.2. The NAO signal

To test the robustness of NAO response further, a principal component analysis (PCA) was performed on the JFM Φ_{1000} anomalies polewards of 20 °N in LABMAX, LABMIN and CTRL. The Φ_{1000} anomalies were then regressed onto the leading principal component (PC) produced by the PCA. The resulting regression pattern is equal to the leading empirical orthogonal function (Thompson and Wallace, 1998) shown in Figure 3. This pattern, which does not differ much from the pattern found when doing PCA on CTRL separately (not shown), will in the following be referred to as the model NAO (NAO_m). The leading PC will similarly be referred to as the NAO_m index. The NAO_m's centres of action in the Arctic and the Atlantic sector are situated slightly eastwards, and are slightly more zonally shaped than in corresponding fields based on observed/reanalysed data in Thompson and Wallace (1998). This is perhaps related to the fact that the model's storm tracks in this region are too zonally confined (Doblas-Reyes *et al.*, 1998; Lopez *et al.*, 2000). The NAO_m index (not shown) indicates that the NAO_m varies similarly from month to month in the CTRL simulation and in LABMAX and LABMIN. Based on the 42 months with data from each of the experiments, we find that the mean JFM NAO_m decreases with increasing sea-ice extent in the Labrador Sea (Figure 4). Figure 4 shows that there is a significant difference between the three experiments. A one-way analysis of variance of the three experiments

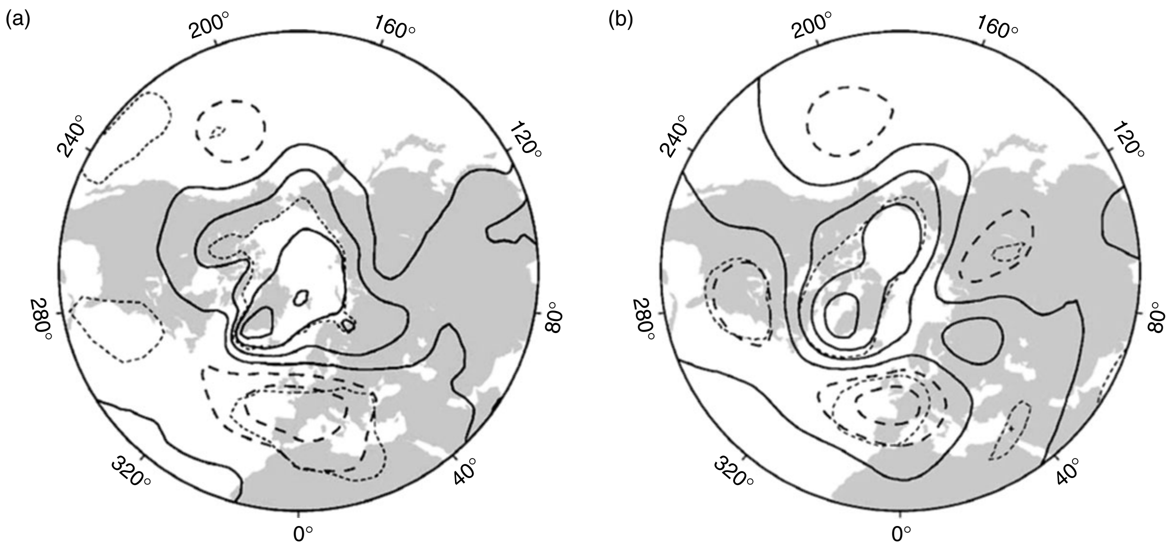


Figure 2. (a) Long-term mean JFM difference, LABMAX − LABMIN, in Φ_{1000}. Contour interval is 10 m, positive (solid) and negative (dashed). The zero contour is omitted. (b) As (a), but for Φ_{500}. Contour interval is 15 m, positive (solid) and negative (dashed). In both (a) and (b) the thin dashed line encircles areas where the difference/response is statistically significant at the 5% level

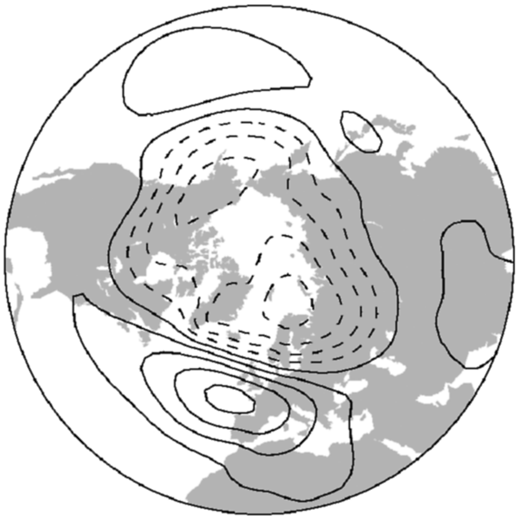


Figure 3. Geopotential height Φ_{1000} anomalies in each of the three winter months (JFM) for all the runs (LABMAX, CTRL, LABMIN) regressed onto the NAO_m index (see text for further explanation). Contour interval is 10 m, positive (solid) and negative (dashed). The zero line is omitted

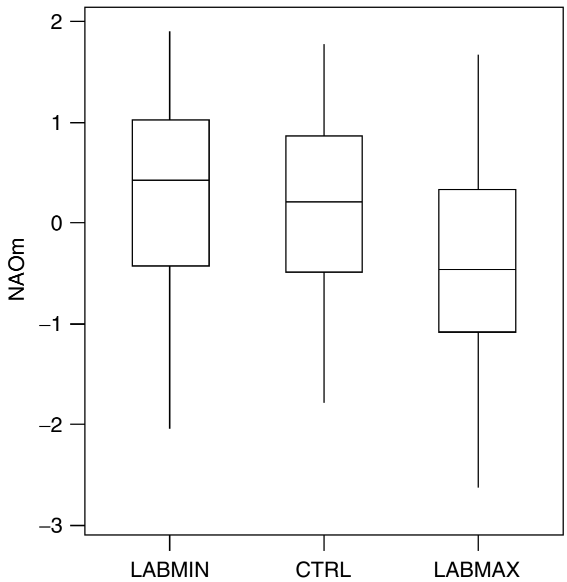


Figure 4. Box plots of winter means of NAO_m standardized indices for all experiments on y-axis versus experiment along the x-axis. The line drawn across each box indicates the median, or middle, of the data. The bottom and top edges of the box mark the first (25th percentile) and third (75th percentile) quartiles respectively. The whiskers go up and down to the maximum values. For each dataset (LABMIN, CTRL and LABMAX) $N = 42$, 3×14 years

shows that the Labrador sea-ice is a statistically significant factor in controlling the mean NAO_m at more than 99% confidence (F-ratio of variance explained by NAO: $= 5.61$; $p = 0.005$).

3.3. Storm-track response

The mean February standard deviation of band-pass-filtered 500 hPa height has been calculated for the 2–10 day (S_{HF}) frequency band, which is associated with synoptic activity (Doblas-Reyes and Deque, 1998).

In Figure 5(a), S_{HF} for the CTRL experiment is shown together with the difference in synoptic activity (ΔS_{HF}) between LABMAX and LABMIN. All three simulations exhibit a maximum across the North Atlantic, quite similar to the climatological S_{HF} (Chang *et al*., 2002). This maximum is interpreted as the North Atlantic storm track. Having positive values in the ΔS_{HF} field to the north and negative to the south of the storm-track belt indicates a northward shift in the North Atlantic storm track (Figure 5(b)). In this context, the shift described is associated with a reduction of the Labrador sea-ice (and an opposite shift would take place for an increase in Labrador sea-ice).

The simulated synoptic activity has also been investigated using a feature-based tracking approach (Hodges, 1994, 1995, 1996, 1999; Hoskins and Hodges, 2001). Figure 6 shows the winter mean difference in cyclone track number density between LABMAX and LABMIN. When there is less ice in the Labrador Sea, this pattern shows that we have a northward shift in the North Atlantic cyclone tracks, slightly tilted in the northwest–southeast direction. This pattern resembles a corresponding difference in number density

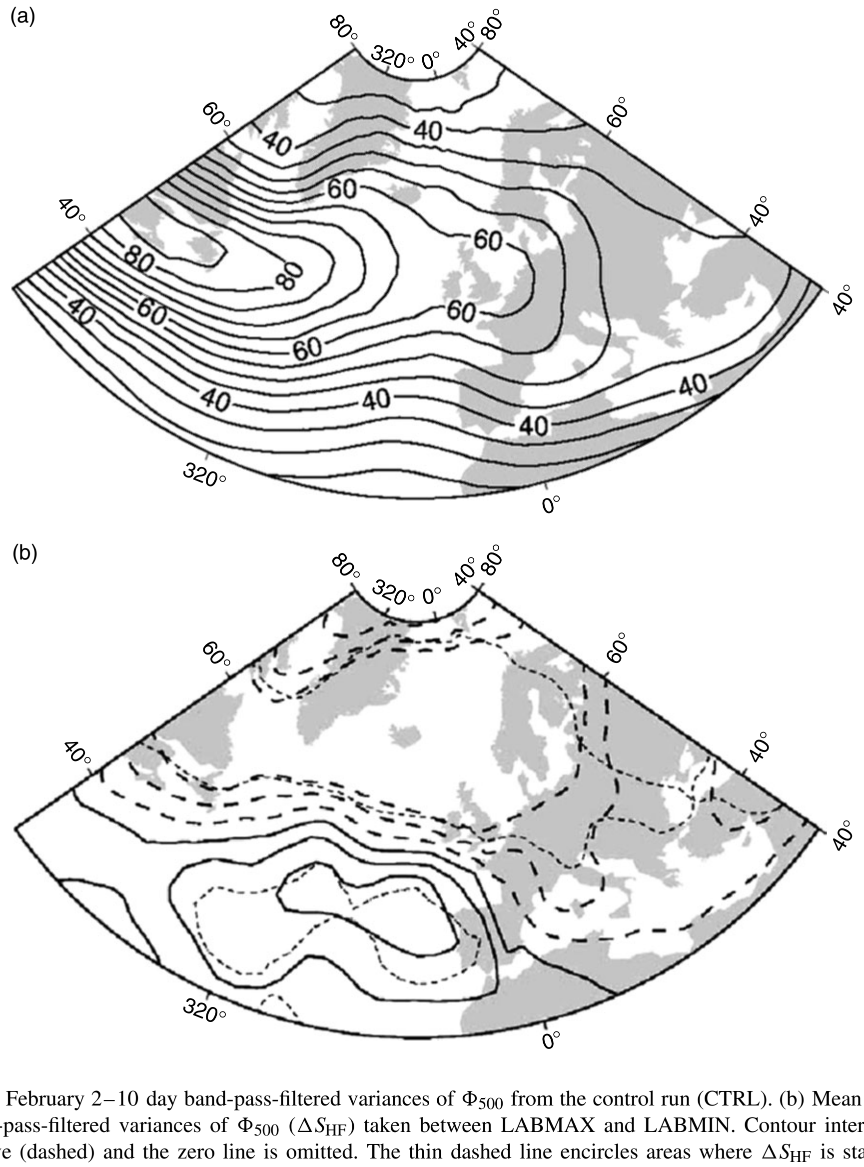
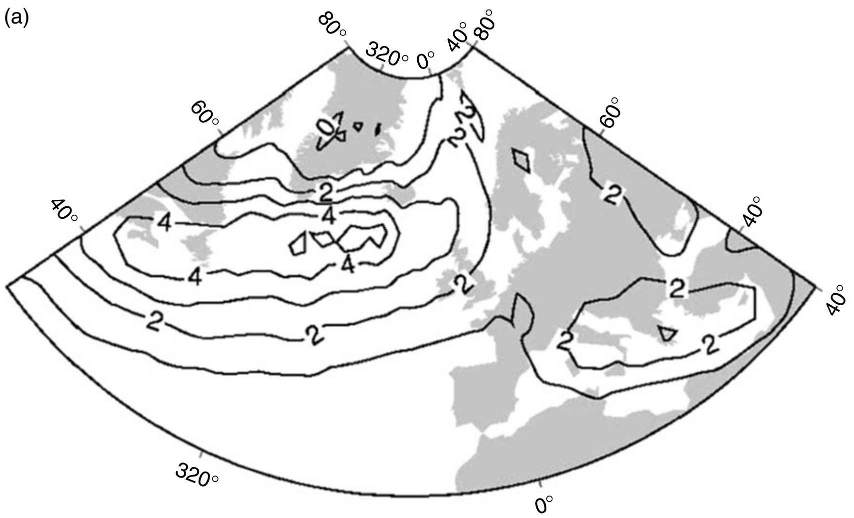


Figure 5. (a) Mean February 2–10 day band-pass-filtered variances of Φ_{500} from the control run (CTRL). (b) Mean February difference in 2–10 day band-pass-filtered variances of Φ_{500} (ΔS_{HF}) taken between LABMAX and LABMIN. Contour interval is 3 m, positive (solid) and negative (dashed) and the zero line is omitted. The thin dashed line encircles areas where ΔS_{HF} is statistically significant at the 5% level



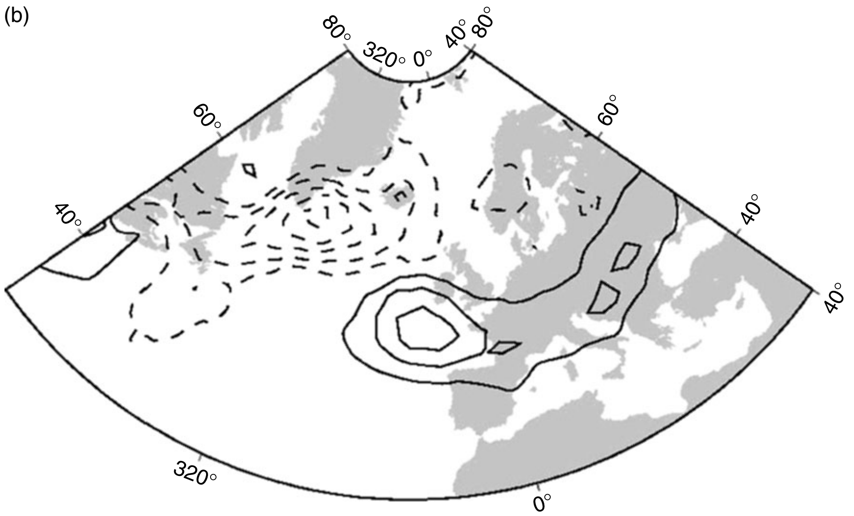


Figure 6. (a) Mean wintertime (JFM) cyclone track density as simulated with the control run (CTRL). Densities have been scaled to number densities per 5° (geodesic radius) spherical cap (approximately 1×10^6 km^2) per winter using the computed probability density function (PDF). (b) Mean difference (LABMAX – LABMIN) in wintertime (JFM) cyclone track density; contour interval is 0.2, positive (solid) and negative (dashed)

of cyclogenesis (start points of individual storm tracks; not shown). The difference in number density of cyclolysis (end points of individual storm tracks) between the two perturbed runs is, however, more scattered (not shown). The most remarkable finding in the cyclolysis response is that fewer cyclones end their life cycle in the Labrador Sea during winters with reduced ice extent (LABMIN). Moving east, on the southeast coast of Greenland, we find the opposite conditions: more cyclolysis events during winters with decreased sea ice in the Labrador Sea.

Reduced sea ice in the Labrador Sea reduces the cyclolysis activity in this area and extends the cyclone routes farther downstream in the lee of southern Greenland, where an increase in cyclogenesis has taken place as well (not shown). In addition, or maybe because of this fact, the majority of the individual North Atlantic cyclone tracks seem to have a more northerly position in the LABMIN case.

Following Hoskins *et al*. (1985), a warm surface anomaly can be considered as a positive potential vorticity anomaly, which can intensify existing cyclones or upper-level potential vorticity anomalies of the same horizontal scale approaching this area. This can cause a downstream extension of the route of the existing cyclones and downstream cyclogenesis due to the baroclinic intensification of the upper-level potential vorticity anomalies. In addition, increased areas of open sea lead to increased latent and sensible heat fluxes into the atmosphere that are known to have a significant impact on cyclone development (Uccellini, 1990; Grønås *et al*., 1994; Grønås, 1995). With respect to the baroclinic and latent heat mechanisms mentioned, a cold SST (and positive sea-ice extent) anomaly will have the opposite effect on the cyclones.

4. DISCUSSION AND CONCLUSIONS

This study has shown that a local sea-ice-extent perturbation in the Labrador Sea can induce a substantial statistically significant NAO–AO response. Increased Labrador sea-ice extent (LABMAX) results in an NAO_m index that is, on average, 0.4 standard deviations below the mean value of the control run (Figure 4). Decreased Labrador sea-ice extent (LABMIN) leads to a mean NAO_m index that is 0.3 standard deviations above the mean (Figure 4). One standard deviation of winter mean NAO_m corresponds to a maximum anomaly in Φ_{1000} of 40 m (Figure 2(a)).

There is little resemblance of our mean response to the response found by Lopez *et al*. (2000). In particular, Lopez *et al*. (2000) did not find a characteristic NAO-like response. Lopez *et al*. (2000), however, investigated the atmospheric response to a number of configurations of a multi-poled SST anomaly, but point out clearly the dominance of the Labrador Sea SSTs on the North Atlantic response. One should not expect a large degree of similarity between the results presented here and the results in Lopez *et al*. (2000), because of the very different scale and shape of the anomalies used in the two studies. Our results agree more with findings in a recent study by Magnusdottir *et al*. (in press), who investigated the simulated atmospheric response to both North Atlantic sea-ice perturbations only and combined SST–sea-ice perturbations. Their perturbations were constructed by integrating the last 40 years' trends in North Atlantic SSTs and sea ice. These anomalies have been amplified (equivalent to an integrated 200 years' trend) and added to (subtracted from) the climatology. The response patterns shown by Magnusdottir *et al*. (in press) associated with Labrador sea-ice (and SST) conditions similar to the ones used in the present study, are strikingly similar to the responses found here. Since SST and sea-ice anomalies in Magnusdottir *et al*. (in press) are multi-poled, involving both the Labrador Sea and Greenland Sea, the results of their study and ours strongly suggest that the Labrador Sea has a prevailing influence on the large-scale North Atlantic circulation. This is in qualitative agreement with Lopez *et al*. (2000). But, as mentioned before, the characteristics of the atmospheric responses are different, and this difference is probably due to the different characteristics of the anomalies employed.

A second feature of our results is that the imposed sea-ice anomaly has a strong impact on the high-frequency synoptic variability. Our diagnostics show that the imposed sea-ice anomaly strongly affects the North Atlantic extratropical cyclones, which help propagate the signal across the basin to create an NAO–AO response. Reduced sea-ice cover in the Labrador Sea causes a northward shift in the North Atlantic cyclone tracks, which is consistent with the increase in NAO_m index. Increased sea-ice cover gave a southward shift in the North Atlantic cyclone tracks, which is also consistent with the overall reduction of the NAO_m index in this case. The cyclone changes are in accordance with theory for growth and decay of baroclinic disturbances (Hoskins *et al*., 1985; Uccellini, 1990; Grønås *et al*., 1994; Grønås, 1995).

It has long been recognized that fluctuations in the NAO and extratropical SSTs are related (Bjerknes, 1964), and there are clear indications that the North Atlantic Ocean SST varies significantly with the overlying atmosphere. The leading mode of SST variability over the North Atlantic during winter consists of a tri-polar pattern with a cold anomaly in the subpolar region, a warm anomaly in the middle latitudes centred off of Cape Hatteras, and a cold subtropical anomaly between the equator and 30 °N (e.g. Deser and Blackmon, 1993; Kushnir, 1994). The emergence of this pattern is consistent with the observed spatial form of the anomalous surface fluxes associated with the NAO pattern (Cayan, 1992). The strength of the correlation increases when the NAO index leads the SST, which indicates that SST is responding to atmospheric forcing on monthly time scales (Battisti *et al*., 1995; Delworth, 1996; Deser and Timlin, 1997). Changes in sea-ice cover in both the Labrador and Greenland Seas, as well as over the Arctic, are well correlated with the NAO (Deser *et al.*, 2000) and are coherent with the SST variations related to the NAO, in the sense that positive SST anomalies correspond to negative sea-ice anomalies and vice versa. The relationship between the sea-level pressure (SLP) and ice anomaly fields is consistent with the notion that atmospheric circulation anomalies force the sea-ice variations (Prisenberg *et al.*, 1997). In our experiment, it is interesting to see that the causal relationship between the NAO phase and Labrador sea-ice extent is opposite to the positive correlation found in observations. However, in this study, the sea-ice change is the imposed forcing and the simulated circulation changes constitute the response. The reversed relationship can thus be interpreted as a negative feedback, in the sense that the atmospheric response would oppose the imposed sea-ice anomaly. Hence, the atmospheric response to sea-ice variations in the Labrador Sea would act as a negative feedback mechanism that attenuates NAO variations. From the experiments performed here it is not possible to estimate on what time scales such a mechanism is most active. However, according to Deser *et al.* (2000), such feedbacks are likely to take place on the longer time scales (decade to century). Since LABMAX produces the strongest response, an interesting future experiment would be to place an even more extensive ice cover in the Labrador Sea and see if this produces an even lower NAO index. Such a result would help shed light on the possible nonlinearity of the response. A more speculative thought is that the Labrador sea-ice may also be relevant for interpreting nonlocal atmospheric responses to rapid changes in fresh water fluxes in this area observed to occur in palaeoclimatic records.

Finally, it should be noted that although the simulated response to the Labrador sea-ice reported here is statistically significant at the 5% level, the result could be model dependent and sensitive, for example, to the boundary-layer flux parameterizations used in the AGCM. It would be of interest to investigate whether the result is reproducible with other AGCMs and investigate the sensitivity of the response to surface flux parameterizations.

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