# Large-scale impact of localised Labrador sea-ice changes on the North Atlantic Oscillation

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## *Abstract*

The wintertime atmospheric response to sea surface temperature and sea-ice anomalies in the Labrador Sea has been investigated by means of an atmospheric general circulation model. The simulated response to low temperatures and heavy ice conditions in the Labrador Sea shows a statistically significant (at 95% confidence) negative North Atlantic Oscillation (NAO) response in the low-level pressure field. Conversely, reduced sea-ice in the Labrador Sea is associated with positive NAO response. This result indicates that atmospheric response to changes in the sea-ice (and sea surface temperatures) in the Labrador Sea can constitute a negative feedback, which may cause long-term phase shifts in NAO. This large-scale response to a local perturbation of sea-ice conditions is most likely brought about by changes in the synoptic eddies. Changes in the sea-ice cover cause marked changes in low level baroclinicity that perturb the travelling baroclinic disturbances, which then bring the signal downstream to manifest a non-local response. The two simulations with opposite sea ice conditions in the Labrador Sea exhibit a maximum mean wintertime difference in the 1000 hPa height field of 40m. It is also found that the changes in sea-ice produce a statistically significant change in the NAO index of 0.7 standard deviations.

KEY WORDS: GCM sensitivity, Labrador sea-ice perturbations, NAO-response

## *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. It also influences the upper ocean stratification and is important for deep-water formation. Alternatively, sea-ice is in turn influenced by both the oceanic and atmospheric circulations. In a recent study, Deser *et al.* (2000) described the Arctic sea-ice extent variability during the past 50 years. They noted that the largest variability occurs in winter over the Atlantic sector. The leading mode of sea-ice variability in wintertime, which accounts for 35% of the total variance, 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 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 GIN Seas. Closer investigations of lead/lag relationships indicate that Arctic sea-ice is responding to atmospheric forcing on monthly to interannual time-scales (Prisenberg *et al.* 1997, Deser *et al.* 2000). On longer timescales, these relationships are more questionable 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).

In general, the number of published atmospheric general circulation model (AGCM) studies of atmospheric response to sea-ice cover anomalies only is much less than the number of SST sensitivity studies with AGCMs (Lopez *et al.* 2000; Kushnir *et al.*

2002). Lopez *et al.* (2000) addressed atmospheric response to changes in North Atlantic sea-ice cover. However, they simulated the atmospheric response to a complex, non-local combined SST- and sea-ice anomaly in the North Atlantic. A series of four perturbation experiments were then carried out with different sign combination of the pattern poles in the SST/sea-ice anomaly. The analysis of the simulated atmospheric responses showed that Labrador Sea surface conditions were the main controlling factor in producing the atmospheric response of the four patterns. However, this statement was not verified in Lopez *et al.* (2000) since a separate experiment with a single localised anomaly in the Labrador Sea was not carried out. This is a necessary, but probably not sufficient task to do in order 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 larger impact on the surface energy fluxes than SST changes in already open sea areas. Furthermore, earlier studies have demonstrated only a weak response to North Atlantic SST anomalies (Kushnir *et al.* 2002; and references therein).

In the present study, we investigate the atmospheric response to a sea-ice cover anomaly in the Labrador Sea. The climatic importance of the Labrador Sea is large (Labrador 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.* (2002) 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 examined time-scales. These relations further underline the importance of quantifying the atmospheric effect of surface conditions in the Labrador Sea.

The plot of this paper is as follows. The following section describes our experimental design. In Section 3 we present the model results. Here we focus on both mean conditions and variability. Large differences between the full AGCM response and the direct linear response to mid-latitude SST anomalies are often due to the eddy response in the full system (Kushnir *et al.* 2002). We therefore offer 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 (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°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 (60W-50W, 55N-60N) 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 March in the period 1979-1995 was constructed. A local 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 with ±3, corresponding to ±3 standard deviations (StD) of the SST index and superposed on the January, February and March climatological SST dataset (Reynolds and Smith 1994) (Figure 1). The SST index spans 5 StD during these 16 analysed winters (80 realisations) and in the area of highest variability, one StD corresponds to 1ºC.

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. From CTRL, 14 initial states from 14 different November months were extracted. For each initial state, two winter runs (Nov-Mar) were made, one with maximal (LABMAX) and one with minimal (LABMIN) Labrador Sea-ice-cover (Figure 1). 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 selected SST pattern was a non-local SST tripole whereas our anomaly is localised over the Labrador region. Compared to 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 took place in the Little Ice Age. That resulted in maximum SST amplitudes of around 5ºC whereas the maximum amplitude in our study is around 3ºC.

Since the large sea-ice variability in the Labrador Sea seems to be a part of a east-west dipole pattern on the monthly to interannual time-scales (Deser *et al.* 2000), isolation of the surface conditions in this area might seem unphysical. However, the pattern in Deser *et al.* 2000 explains 35% of the sea-ice variance and a considerable amount of variability might therefore be explainable by this localised 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 timescales may be of particular relevance here since the atmospheric feedback is likely to take place on these timescales (Deser *et al.* 2000).

## *3. Results*

### 3.1 Mean Response

The winter mean difference between LABMAX and LABMIN in geopotential height at 1000 hPa ($\Phi_{1000}$) is shown in Figure 2a. The corresponding difference between each of the perturbed experiments and CTRL are not shown. However, these responses have a similar pattern as the field in Fig. 2a, but with weaker amplitude (and opposite sign in the LABMIN-CTRL case). It can be seen in Figure 2b that the response is nearly equivalent barotropic since the pattern of the response in $\Phi_{500}$ is similar to the $\Phi_{1000}$ response. The patterns bear striking resemblance to the NAO/AO pattern (Hurrell 1995, Thompson and Wallace 1998, Ambaum *et al.* 2001). Since a two-sided Student t-test gave that substantial areas of the response fields are statistically significant (at the 95% level), our results suggest that a local sea-ice anomaly in the Labrador Sea has created a statistically significant NAO-like response.

### 3.2 The NAO signal

To further test the robustness of NAO response, a principal component analysis (PCA) has been performed on the JFM $\Phi_{1000}$ anomalies polewards of 20°N in LABMAX, LABMIN and CTRL. The $\Phi_{1000}$ 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 (EOF) (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 stormtracks 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 (ANOVA) of the three experiments shows that the Labrador sea-ice is a statistically significant factor in controlling the mean $NAO_m$ at more than 99% confidence (p-value = 0.005).

### 3.3 Storm track response

The mean January to February StD of band-pass filtered 500hPa height has been calculated for the 2-10 day ($S_{HF}$) frequency band, which is associated with synoptic activity (Doblas-Reyes and Deque 1998). The reason for including only January and February is the large filter width, which is 30 days. In Figure 5a $S_{HF}$ for the CTRL experiment is shown together with the difference in synoptic activity ($\Delta S_{HF}$) between LABMAX and LABMIN. All the three simulations exhibit a maximum across the North Atlantic, quite similar to the climatological $S_{HF}$. This maximum is interpreted as the North Atlantic storm-track. Having positive values in the $S_{HF}$ 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 5a). In this context the described shift is associated with a reduction of the Labrador sea-ice (and an opposite shift would take place for an increase in Labrador sea-ice).

Further investigation of the simulated synoptic activity in a Lagrangian framework (Hodges 1994; 1995; 1996; 1999; Hoskins and Hodges 2001) has been performed in order to get a more detailed picture of the eddy response. In Figure 6, 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 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 Labrador Sea ice.

Removal of sea-ice in the Labrador Sea reduces the cyclolysis activity in this area and extends the cyclones' route farther downstream in the lee of southern Greenland were 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 may result in a downstream extension of the route of the existing cyclones or downstream cyclogenesis due to the baroclinic intensification of the upper level potential vorticity anomalies. Adding areas of increased open sea leads to increased latent and sensible heat fluxes into the atmosphere. Increased latent heat content is known to have significant impacts on cyclone developments (Uccellini 1990), such as re-intensification (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. Conclusions

This study has shown that a local sea-ice perturbation in the Labrador Sea can induce a statistically significant NAO-like response, where high winter sea-ice cover (LABMAX) results in a $NAO_m$ index, which on average is 0.4 StD below the mean value (which is zero) of the control run. Simulation of lower than normal Labrador sea-ice extent (LABMIN) leads to a mean $NAO_m$ index which is 0.3 StD above zero. One StD corresponds to a maximum anomaly in $\Phi_{1000}$ of 40m.

There is little resemblance of our mean response with similar diagnostics shown in Lopez *et al.* (2000). In particular Lopez *et al.* (2000) did not find a characteristic NAO-like response. They have, however, investigated the atmospheric response to a number of configurations of a multi-poled SST anomaly, but point clearly out 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 another study by Magnusdottir *et al.* (2003). Magnusdottir *et al.* (2003) have investigated simulated atmospheric response to both North Atlantic sea-ice only- and combined SST/sea-ice perturbations. Their perturbations have been 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 (subtracted) to (from) climatology. The responses shown in Magnusdottir *et al.* (2003) associated with Labrador sea-ice (and SST) conditions similar to the ones used in the present study, are strikingly similar to the responses presented above. This is the case for both the mean and NAO response. Since SST and sea-ice anomalies in Magnusdottir *et al.* (2003) are multi-poled, involving both the Labrador Sea and Greenland Sea, the combined results of the two studies 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 in the anomalies employed.

A second feature of the results presented above is that the imposed sea-ice anomaly has strongest impact on the high frequency synoptic variability. Our diagnostics suggest that the imposed sea-ice anomaly affects the travelling cyclones which brings the signal downstream and thus creates a remote response. The effect of a reduced sea-ice cover in the Labrador Sea is a northward shift in the North Atlantic cyclone tracks, which is consistent with the increase in $NAO_m$ index. Heavy sea-ice conditions gave a southward shift in the North Atlantic cyclone tracks, which also is consistent with the overall reduction of the $NAO_m$ index in this case. The diagnostics in this study show that the described cyclone track behaviour is in accordance with theory for growth and decay of baroclinic disturbances (Hoskins *et al.* 1985; Uccellini 1990).

One should note that it has long been recognized that fluctuations in SST and the strength of the NAO are related (Bjerknes 1964), and there are clear indications that the

North Atlantic Ocean 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 spatial form of the anomalous surface fluxes associated with the NAO pattern (Cayan 1992b). 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 appear to be well correlated with the NAO (Deser *et al.* 2000) and they 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 in the Labrador Sea, the relationship between the NAO phase and sea-ice/SST is reversed compared to what has been observed. 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 could be a negative feedback mechanism that may attenuate NAO variations. From the experiments performed here it is not possible to estimate on which time-scales such a mechanism is most active. However, according to Deser at al (2000) they are likely to take place on the longer time-scales (decade to century). Since LABMAX seems to produce 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 add confidence to the present findings. A more speculative thought is that it might also be relevant for interpreting non-local atmospheric responses to observed rapid changes in fresh water fluxes in this area found in paleoclimatic records.

Finally, it should be noted that although the simulated response to the Labrador Sea SST/sea-ice anomaly as reported upon here is robust, the result could be model dependent and sensitive, for example, to the boundary layer flux parameterizations used in the AGCM. It would consequently be of interest to examine whether the result is reproducible with other AGCMs.

## 5. Acknowledgments

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## *Figure captions*

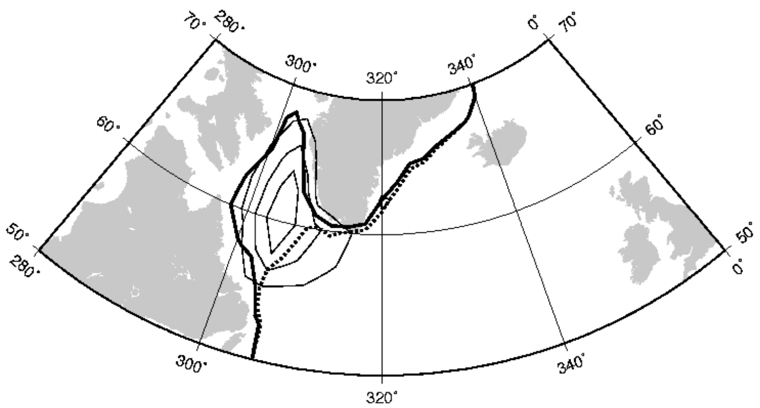


Figure 1. Thin solid contours: Labrador Sea SST anomaly. Contour interval is 1K. Thick solid contour shows the perturbed sea-ice border for January when the SST anomaly has been subtracted from climatology (LABMAX). Thick dotted line shows perturbed sea-ice boundary for January when the SST anomaly has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

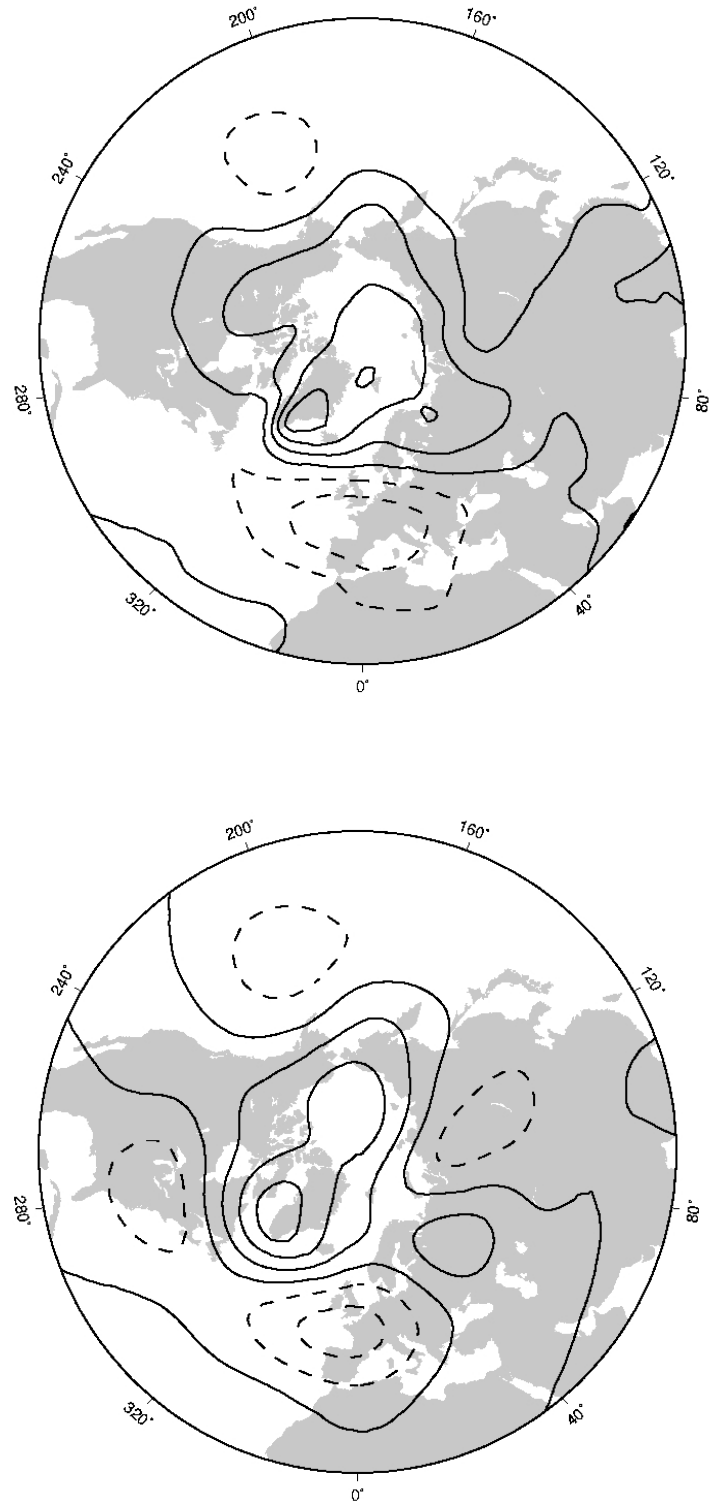


Figure 2. a) Long term mean JFM difference, LABMAX - LABMIN, in $\Phi_{1000}$. Contour interval is 10m, positive (solid) and negative (dashed). The zero contour is omitted. b) Same as in a, but for $\Phi_{500}$ . Contour interval is 15m, positive (solid) and negative (dashed).

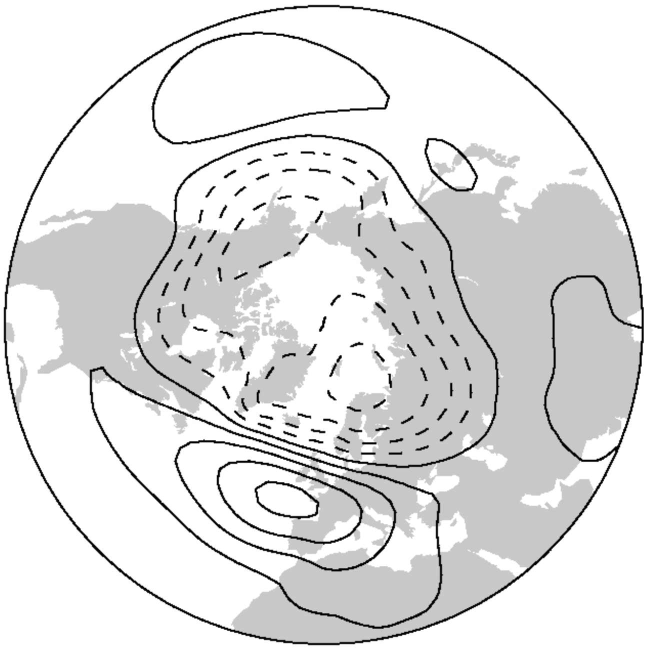

Figure 3. $\Phi_{1000}$ anomalies in each of the three winter months (JFM) for all the runs (LABMAX, CTRL, LABMIN) regressed onto the $NAO_m$ index (see text for further explanation). Contour interval is 10m, positive (solid) and negative (dashed). The zero line is omitted.

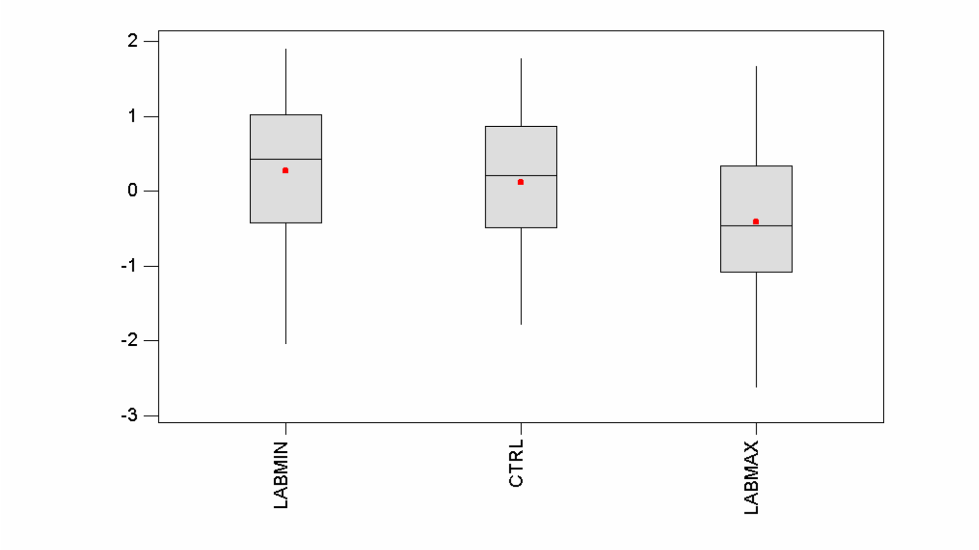


Figure 4. Boxplot of $NAO_m$ (in standard deviations) on y-axis versus experiment along the x-axis. The line drawn across each box indicates the median, or middle, of the data. The bottom and top edges of the box mark the first ($25^{th}$ percentile) and third ($75^{th}$ percentile) quartiles respectively. The red dots mark the mean value of $NAO_m$ for each experiment.

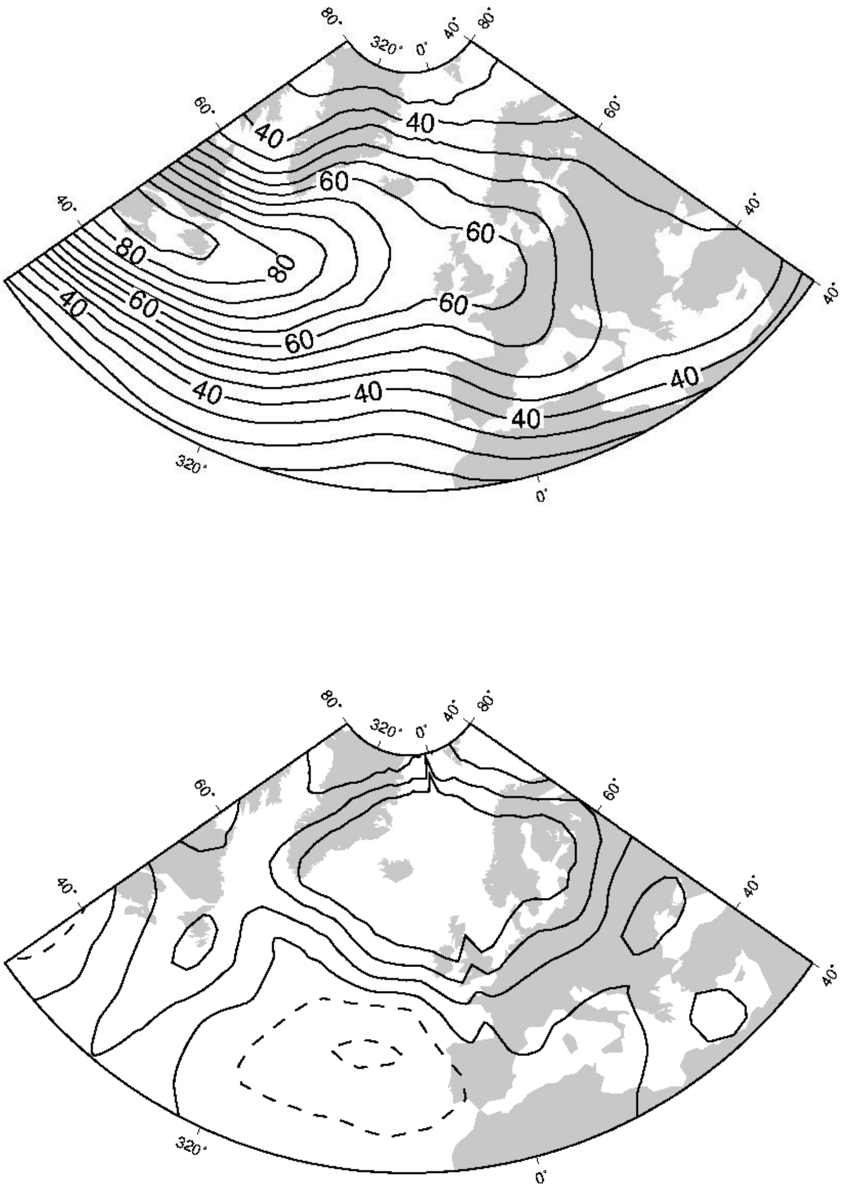


Figure 5. a) Mean JF 2-10 day band pass filtered variances of $\Phi_{500}$ from the control run (CTRL). b) Mean JF difference in 2-10 day band-pass filtered variances of $\Phi_{500}$ taken between LABMAX and LABMIN. Contour interval is 3m, positive (solid) and negative (dashed). The zero line is omitted.

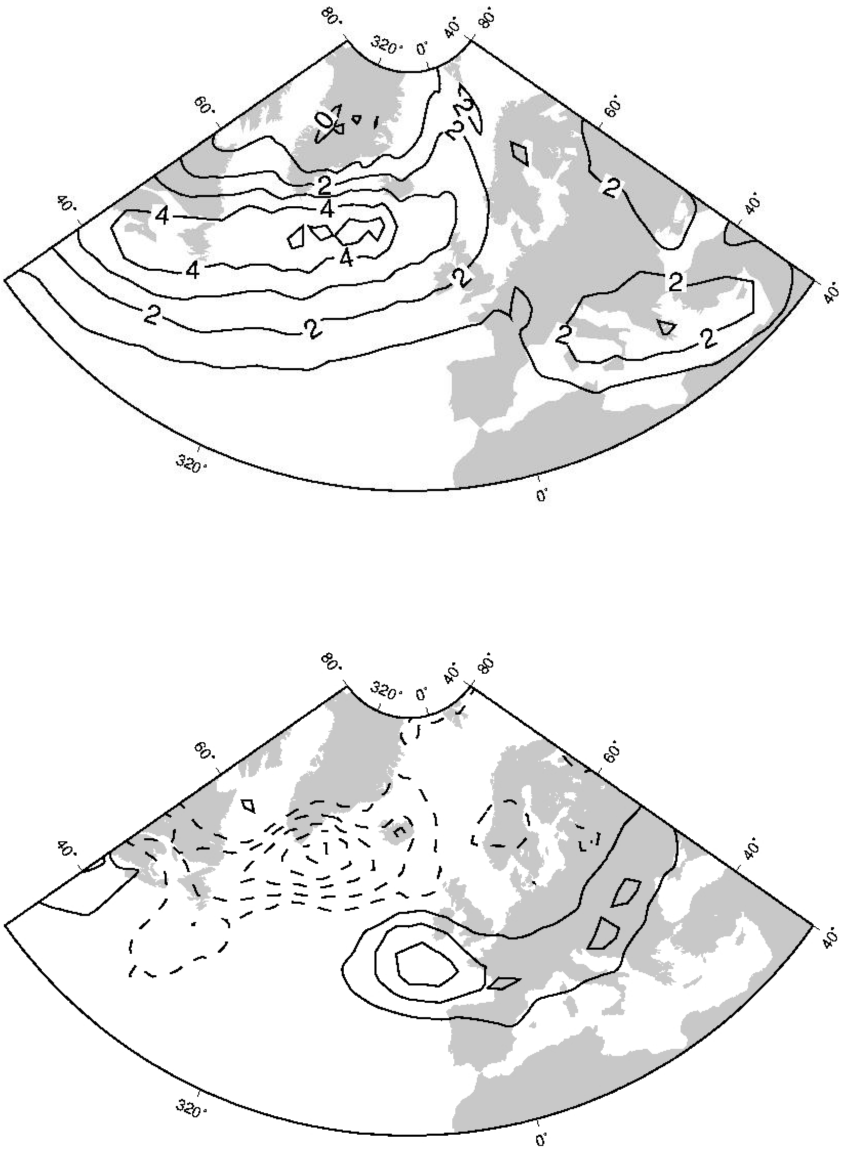


Figure 6. a) Mean wintertime (DJFM) cyclone track density as simulated with the control run (CTRL). Densities have been scaled to number densities per 5° (geodesic radius) spherical cap (approximately $1*10^6 km^2$) per winter using the computed probability density function (PDF). b) Mean difference,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