The North Atlantic Oscillation–Arctic Oscillation in the CCSM2 and Its Influence on Arctic Climate Variability

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ABSTRACT
Recent observations suggest that large and widespread changes are occurring in the Arctic climate system. Many of these are associated with the North Atlantic Oscillation (NAO) or the closely related Arctic Oscillation (AO). Here, the Arctic climate and its response to the NAO–AO is examined in a control simulation of the newly released Community Climate System Model, version 2 (CCSM2). Variability in the atmosphere and sea ice systems are considered and the physical mechanisms that drive the variations are discussed. It is found that the model reasonably simulates the spatial structure and variance of the sea level pressure, surface air temperature, and precipitation associated with the NAO–AO. The sea ice response to the NAO–AO also compares well to observations. However, it varies over the length of the time series, which is related to variations in the spatial structure of the sea level pressure anomalies associated with the NAO–AO over time. The model results suggest that these variations, which are similar to changes that occur over the observed record, are common and part of the natural variability of the system. However, the magnitude of the observed trends over the last 40 yr in the NAO–AO index are never realized in the model simulations, suggesting that these trends may be associated with changes in anthropogenic forcing, which the simulation examined here does not include.

1. Introduction and motivation
The Arctic climate system has undergone substantial changes in recent years. These are widespread and are present in the atmospheric, oceanic, sea ice, terrestrial, and biological systems (e.g., Serreze et al. 2000). They include, among other changes, a decrease in Arctic atmospheric sea level pressure, a warming of the waters of Atlantic origin within the Arctic Ocean, a decrease in summer sea ice cover, and a thawing of the permafrost. Many of these changes appear to be interrelated and to be associated with variability and trends in the North Atlantic Oscillation (NAO) or the closely related Arctic Oscillation (AO; Thompson and Wallace 1998).

The North Atlantic Oscillation is the dominant mode of variability in the North Atlantic region and represents a redistribution of atmospheric mass between centers of action located near the Azores high and the Icelandic low. A high NAO phase indicates a strengthening of the Azores high and the Icelandic low, increasing the sea level pressure (SLP) gradient between these two centers, and resulting in stronger westerly flow. This has associated changes in surface air temperature (SAT) and precipitation (Hurrell 1995, 1996). The observed NAO index has shown a substantial upward trend over the last 40 years, which is unprecedented over the instrumental record. The NAO trend has related trends in atmosphere, ocean, and sea ice conditions as discussed below.

The Arctic Oscillation (also referred to as the northern annular mode) is closely related to the NAO, with the observed time series correlating at 0.95 for monthly data (Deser 2000). The AO is the dominant mode of variability in the Northern Hemisphere and exhibits an annular pattern with decreased SLP over the Arctic basin associated with increase SLP at midlatitudes with centers of action in the North Atlantic and North Pacific. As such, the AO essentially encompasses the NAO variability but emphasizes the zonally symmetric nature of the hemispheric variability. There has been some recent debate over the physical relevance of the annular structure and in particular, of the Pacific center of action, of this mode (e.g., Deser 2000; Ambaum et al. 2001; Wallace and Thompson 2002). This debate is beyond the scope of this paper and due to the high correlation between the NAO and AO, we will use them interchangeably and refer to the leading mode of variability as the NAO–AO.

The NAO–AO has been associated with climate conditions in the high-latitude North Atlantic and Arctic. There have been a number of studies detailing this influence. Here the focus is on those studies that concern Arctic sea ice conditions. Previous studies have found...
that the leading winter sea ice variability pattern consists of a dipole with opposing centers of action in the Davis Strait–Labrador Sea and the Greenland and Barents Seas (e.g., Walsh and Johnson 1979; Rogers and van Loon 1979; Fang and Wallace 1994; Deser et al. 2000). These ice conditions are related to the temporal variability of the NAO–AO, with a high NAO–AO index associated with extensive Labrador Sea ice cover and reduced Greenland and Barents Sea ice cover. The atmospheric conditions lead the sea ice anomalies (e.g., Walsh and Sater 1981; Fang and Wallace 1994), indicating that the atmosphere is forcing the ice anomalies, probably through a combination of dynamic and thermodynamic effects.

The summertime sea ice area shows a decreasing trend of 0.6% yr$^{-1}$ over the last 20 yr (e.g., Maslanik et al. 1996) with the maximum reduction occurring in the Siberian Sea (Parkinson et al. 1999). This decrease is likely related to a decrease in sea level pressure over the central Arctic (e.g., Walsh et al. 1996), which appears to be influenced in part by the NAO–AO (Serreze et al. 1997). Indeed, Deser et al. (2000) show that the Siberian Sea ice extent variations are initiated by spring SLP anomalies that are related to the winter NAO–AO index. However, they also note that there is not a consistent relationship between the winter NAO–AO and spring SLP anomalies over the high Arctic. Additionally, as discussed by Dickson et al. (2000), the influence of the NAO on storms and the storm-track “tapers off rapidly toward the high Arctic,” suggesting that there are other factors that contribute to the Arctic SLP anomalies.

Dramatic decreasing trends in ice thickness are also apparent in the observed record (Rothrock et al. 1999). However, a recent paper by Holloway and Sou (2002) cautions that some of this may be due to inadequate spatial and temporal sampling and suggests that a redistribution of ice mass within the Arctic is more likely. This redistribution has also been noted in other studies. For example, Zhang et al. (2000) use hindcast modeling studies to argue that from the period 1979–88 to the period 1988–96, the eastern Arctic ice volume has decreased by 25% while the western Arctic ice volume has increased by 16%. This is related to changes in the wind-forced ice transport. The recent changes in ice motion and volume appear to be consistent with forcing by the NAO–AO (Rigor et al. 2002). Additionally, changes in ice motion over the observed record have influenced the ice transport through Fram Strait, with increased ice export being linked to a positive NAO index (Kwok and Rothrock 1999). However, as noted by Hilmer and Jung (1999), using a hindcast modeling study, the relationship between the NAO and ice export has changed considerably over the last 40 yr due to a shift in the spatial SLP pattern associated with the NAO.

The previously mentioned studies suggest that the NAO–AO has an important influence on Arctic climate. However, the physical mechanisms by which this occurs are not always clear. For example, the interplay between dynamic and thermodynamic forcing of sea ice conditions is difficult to assess from observations. Additionally, the recent upward trend in the NAO–AO dominates the observed time series and probably the Arctic response. Thus, our understanding of the variability of the Arctic system is biased by this trend.

It remains an active research topic to determine whether the NAO–AO trend is part of the natural variability of the system or is anthropogenically forced. In previous work, a number of studies have examined the natural variability of the NAO–AO and/or its response to changing anthropogenic forcing in climate models (e.g., Fyfe et al. 1999; Shindell et al. 1999; Osborn et al. 1999; Paeth et al. 1999; Robertson 2001; Gillett et al. 2002a). Although there are some discrepancies between individual models, these studies have generally found that the observed NAO–AO trend is unusual compared to control integrations with no changes in greenhouse gases. When changing anthropogenic forcing is applied, most models obtain a significant trend in the simulated NAO–AO. However, these trends are typically weaker than the observed. There has been some debate about whether including a well-resolved stratosphere is necessary to obtain the observed trends, but there is no clear agreement among different studies (e.g., Shindell et al. 1999; Gillett et al. 2002b). The use of model integrations to examine these issues is useful because of the relatively long, complete, self-consistent time series that they produce. Although there are model biases that must be considered in this analysis, the simulated results can provide insight into the mechanisms forcing the system and the likelihood of trends within the context of the model variability.

There are three primary motivations for this study. First, we wish to document the simulation of the North Atlantic Oscillation–Arctic Oscillation and high-latitude climate variability by the Community Climate System Model, version 2 (CCSM2). Second, we want to address the physical mechanisms that drive the simulated variability and links between different components of the climate system. Assuming this variability is realistic, this will give insight into how these processes are occurring in the real climate system. And finally, we wish to assess whether the observed changes in the NAO–AO index and spatial structure are remarkable in the context of the model variability. More specifically, does the model, which does not have changing anthropogenic forcing, attain the same trends seen in the observations? If not, and the simulated variability is reasonable, the suggestion is that anthropogenic forcing is needed to explain recent NAO–AO trends. Alternatively, the model may be missing or inaccurately representing physics that is necessary to reasonably simulate the natural variability of the NAO–AO.

2. Model description

For this study, results from a control integration of the Community Climate System Model, version 2
(CCSM2; J. T. Kiehl and P. R. Gent 2003, personal communication) are examined. This integration is run under present-day conditions with no changes in anthropogenic forcing; 550 yr of model integration are analyzed (years 350–900). This time period was chosen because many of the initial climate drifts in the ice and ocean are considerably reduced by year 350.

CCSM2 is a state-of-the-art coupled general circulation model (GCM) that includes atmosphere, ocean, land, and sea ice components. The model has changed significantly from its initial version (Boville and Gent 1998), with the sea ice and land components being completely modified. This has led to considerable improvements in the polar regions.

The community land model (Bonan et al. 2002) is the land surface component used within the CCSM2. The model includes a subgrid mosaic of land cover types and plant functional types derived from satellite observations, a 10-layer soil model that explicitly treats liquid water and ice, a multilayer snowpack model and a river routing scheme.

The ocean component of the CCSM2 utilizes the Parallel Ocean Program (POP) with a number of improvements (Smith and Gent 2002). In particular, the model uses anisotropic horizontal viscosity, an eddy mixing parameterization (Gent and McWilliams 1990), the K-profile parameterization for vertical mixing (Large et al. 1994), and a more accurate equation of state. The equations of motion are formulated and discretized to allow the use of any locally orthogonal horizontal grid. As such, both the ocean and sea ice systems use a grid that smoothly displaces the model North Pole into Greenland, avoiding problems related to converging meridians in the Arctic. Hence, no filtering of the ocean solution is needed in the Arctic region. The horizontal resolution averages less than 1° and the Bering Strait and Canadian Archipelago are open, allowing for realistic oceanic exchange. As a result of these improvements and the application of river inputs, the Arctic halocline is maintained and the Arctic sea surface salinity compares well to the observed polar hydrographic climatology (Steele et al. 2001).

The community sea ice model incorporated into CCSM2 is a new dynamic–thermodynamic scheme that includes a subgrid-scale ice thickness distribution (Bitz et al. 2001; Lipscomb 2001). The model uses the energy conserving thermodynamics of Bitz and Lipscomb (1999), which has multiple vertical layers and accounts for the thermodynamic influences of brine pockets within the ice cover. The ice dynamics utilizes the elastic-viscous-plastic rheology of Hunke and Dukowicz (1997) with a number of updates. Five ice thickness categories are included within each grid cell and subgrid-scale ridging and rafting of sea ice is parameterized following Rothrock (1975) and Thorndike et al. (1975). A reasonable Arctic sea ice simulation is present in the control integration. The ice area compares very well to observations. However, the ice is relatively thin, which is consistent with biases in the atmospheric radiative forcing at high latitudes (B. Briegleb et al. 2002, personal communication, hereafter BRI).

The atmospheric component of the CCSM2 is the Community Atmosphere Model (CAM2). It builds on the National Center for Atmospheric Research (NCAR) Community Climate Model 3 (CCM3; Kiehl et al. 1996) with a number of improvements and updates. The model has enhanced resolution in the vertical, going from 18 to 26 levels. Other physics improvements include incorporation of a prognostic formulation for cloud water, a generalized geometrical cloud overlap scheme, more accurate treatment of longwave absorption/emission by water vapor, and enhancements to the parameterization of deep cumulus convection. The model is generally run at T42 (~2.875°) resolution.

3. Simulation results

a. Simulated NAO–AO structure and associated atmospheric anomalies

The names and locations of regions used in this study are shown in Fig. 1. Figure 2a shows the regression of winter seasonal (NDJFM) sea level pressure on the principal component of the first empirical orthogonal function (EOF) of sea level pressure in an Arctic–North Atlantic region from the CCSM2 over 550 yr of the integration. The regression is shown to indicate not only the spatial structure in the region of the EOF analysis, but also the teleconnection of this pattern to the Pacific. The first EOF accounts for 47% of the variance. The equivalent analysis from the NCEP–NCAR reanalysis (Kalnay et al. 1996) from 1950–97 using detrended SLP...
data is shown in Fig. 2b for comparison. The EOF from the observational analysis accounts for 41% of the variance. Analysis using a Northern Hemisphere domain, more representative of the Arctic Oscillation, shows very similar characteristics. The principal components associated with the hemispheric and the more regional analysis shown in Fig. 2 correlate at 0.95 for the model results and 0.98 for the NCEP reanalysis. As such, the relationship between the regional or hemispheric EOFs and other climate conditions is nearly identical. Because of this similarity, in the remainder of this paper, the leading mode of variability shown in Fig. 2 will be referred to as the NAO–AO and the term NAO–AO index will refer to the principal component time series from the regional EOF analysis.

The model has a very reasonable simulation of the NAO–AO spatial structure. The primary discrepancy between the simulated and observed NAO–AO is that the simulated variability has a stronger teleconnection with the Pacific. This is similar to results from the Hadley Centre’s coupled model as documented by Osborn et al. (1999). There are other small differences between the model and observations; for example, the Azores high variability is weaker in the model simulations than in the NCEP reanalysis. Additionally, the anomalous low pressure over the Greenland–Iceland–Norwegian (GIN) Seas has a more meridional orientation in the simulation. However, as discussed by Hilmer and Jung (1999), the centers of variability associated with the NAO shift spatially over the observed record and their analysis of data from 1958 to 1977 more strongly resembles the model results.

There is considerable variability in the spatial structure of the simulated NAO–AO. This is illustrated by Fig. 3, which shows the location of the minimum and maximum SLP anomalies associated with the NAO–AO index for 11 50-yr segments of the total 550-yr simulated time series. The high pressure center occurs over a fairly narrow latitude range from 40° to 46°N but varies from 25°W to 14°E longitude. In five of the 50-yr time series, the low pressure center is located over Greenland at 68°N. However, in the remaining six 50-yr segments, the low pressure center is located considerably farther north over the Barents Sea, Kara Sea, and even into the Laptev Sea. It is worth noting that the exact center of this low pressure center can be ambiguous given the somewhat arbitrary pressure reductions to sea level over Greenland. The changes in the spatial structure of the NAO–AO are part of the natural variability of the control simulation examined here, suggesting that this spatial variability may also be a natural part of the real climate system. This result is supported by the analysis of Cassou et al. (2002, manuscript submitted to J. Climate), which shows that a spatial shift in the observed NAO centers of action occurs based on whether the NAO is positive or negative. Alternatively, as discussed by Fyfe (2003), it is possible that the mode of variability is unchanged but is superimposed upon a changing mean state.

It is unclear whether there is an underlying physical cause for the spatial variations in the simulated NAO–AO pattern or whether they result from sampling variability. However, these results do suggest that in 50-yr
realizations of the simulated climate, there can be changes in the spatial structure of the leading mode of SLP variability. These changes can affect how this leading mode relates to other aspects of the climate system, particularly in the Arctic, which is strongly influenced by ocean and sea ice exchanges through narrow straits such as the Fram Strait and the Canadian Arctic Archipelago. Previous climate modeling studies have suggested that variability in the NAO–AO spatial pattern results from anthropogenic forcing. For example, using simulations from a coupled GCM, Ulbrich and Christoph (1999) showed that a northeastern shift in the NAO low pressure center occurred under increasing anthropogenic forcing. While the analysis discussed here does not contradict this result, it does show that changes in the NAO–AO spatial structure can also occur in unforced climate model simulations.

Based on spectral analysis (not shown) the temporal variability of the NAO–AO is also reasonably simulated compared to the observed Hurrell (1995) two-station index from 1864 to 1963. However, over the last 40–50 yr the observed NAO–AO index has been generally increasing. The trends associated with this are unprecedented in the instrumental record (Hurrell 1995). As discussed by Fyfe (2003), this NAO–AO trend may in fact reflect the projection of a forced SLP trend onto an unforced and conservative mode of variability. Figure 4a shows the 40-yr running trend of the simulated NAO–AO index. The trend associated with the observed NAO–AO over the last 40 years is never realized by the model. This is further emphasized in Fig. 4b, which shows the percent of time that a trend of a certain size and length is obtained in the model simulation. The model’s natural climate variability obtains trends equal to or greater than the maximum observed 20-yr trend only 2% of the time. This agrees with previous studies in that the recent observed trend in the NAO–AO appears unusual compared to climate model control integrations. Assuming the model variability is realistic, this suggests that external forcing not present in the model simulation, such as rising atmospheric CO₂ levels, is necessary to explain the recent NAO–AO trends. Further model simulations, with observed external forcing, are needed to investigate this hypothesis. However, while a number of climate models do show significant NAO–AO trends with external forcing, these are typically smaller than the observed trends (e.g., Gillett et al. 2002a). Thus, it is very possible that the model var-
iability may be deficient due to poorly represented processes or missing physics.

The NAO–AO has air temperature and precipitation anomalies associated with it as discussed above. The regression of air temperature on the NAO–AO index for the model simulation is shown in Fig. 5a. As seen in the observations, a significant warming of northern Europe extending into Eurasia is present during a high NAO–AO index. The magnitude of this warming also agrees well with the NCEP reanalysis data (Fig. 5b) and is consistent with previous observational studies (Hurrell 1995; Thompson and Wallace 1998). As present in the observations, a cooling over the Labrador Sea extending into the northwestern Atlantic is also associated with the simulated NAO–AO. There is also a cold signal over Alaska, which is weaker in the observations. The stronger simulated signal is likely due to the relatively strong Pacific teleconnection as discussed above. Analysis using the NCEP reanalysis data also show large temperature changes in the Fram Strait region associated with the NAO–AO. These are not present in the model analysis. However, the large temperature response in these regions may be an artifact of the surface boundary conditions used for the reanalysis that specify either 0% or 100% ice cover per grid cell and likely overestimate some sea ice feedbacks on surface air temperature.

There have been considerable trends in air temperature over high-latitude land cover over the last several decades (e.g., Jones et al. 1999). As discussed by Hurrell (1996), the air temperature trends over northern Europe and Eurasia are related in part to the increasing NAO–AO index. Changes in greenhouse gases also likely contribute to the observed temperature trends. As the simulated NAO–AO rarely obtains the observed NAO–AO trends and no changes in greenhouse gases are present, it is expected that the trends in air temperature in northern Europe and Eurasia would also be lower than recent observations. In fact, the model results indicate that the maximum observed 40-yr air temperature trend in this region (shown in Fig. 5a) occurs less than 1% of the time in the simulation.

The simulated precipitation anomalies associated with the NAO–AO (Fig. 6) also show a considerable similarity to the observations both in their spatial pattern and magnitude. There is increased precipitation over the GIN Seas and Scandinavia, which extends westward across the Atlantic. Reduced precipitation is present over southern Europe. The magnitude of these precipitation anomalies agrees well with the observations of Xie and Arkin (1996) from 1979 to 1995 (Fig. 6b). However, the simulated spatial pattern has the strongest negative anomalies located almost zonally from southern Europe to North America, while the observations have a hint of this but show the larger anomalies over the European continent. This is perhaps related to biases in the simulated climatological precipitation or differences in the spatial pattern of the NAO–AO with the observations for this time period. As shown by Dickson et al. (2000), there is also enhanced precipitation over the Arctic and Arctic drainage basins, suggesting that river flow into the Arctic may be enhanced under high NAO–AO conditions. In fact, the model results show that the river outflow from the Ob, Yenesei, and Lena

Fig. 5. The (a) simulated and (b) observed winter (JFM) surface air temperature regressed on the NAO–AO index. The contour interval is 0.5°C, negative values are shaded, and the zero contour is not shown. The observational analysis uses data from NCEP reanalysis and an NAO–AO index from detrended SLP data. Temperature trends discussed in the text are computed for the Eurasian region shown in (a).
drainage basins is significantly correlated to the simulated NAO–AO index at 0.44.

b. Sea ice anomalies and their association with the NAO–AO

1) Winter variability

It is well documented that the observed winter ice concentration variability exhibits a dipole pattern with enhanced ice cover in the Labrador Sea and reduced coverage in the Greenland–Barents Sea (e.g., Walsh and Johnson 1979; Rogers and van Loon 1979; Fang and Wallace 1994; Deser et al. 2000). This pattern appears to be forced by atmospheric conditions in SLP and air temperature that are associated with the NAO–AO.

The model simulation shows a similar leading mode of wintertime ice variability. Figure 7 shows the first EOF of JFM sea ice concentration from the model simulations and observations. The simulated EOF accounts for 21% of the variance. It has a similar dipole pattern to that seen in the observations and the magnitudes are reasonable. However, the model results reflect some of the biases in the mean state of the ice cover (BRI). In particular, because the mean Labrador Sea ice cover is too extensive, the anomalies are not confined to the shelves but extend over the Labrador Sea. In contrast, the simulated Greenland Sea anomalies are more narrowly confined than the observations suggest. Additionally, the Barents Sea anomalies are reduced compared to the observations and occur farther north due to a more northerly ice edge in the simulations.
The SLP anomalies associated with the leading sea ice EOF are shown in Fig. 8. They bear a strong resemblance to the NAO–AO pattern particularly in the northern North Atlantic–Nordic Sea regions. The principal component of the sea ice EOF and the NAO–AO index are correlated at 0.57, which is similar to the correlation of 0.63 found by Deser et al. (2000) in an analysis of the observed conditions. Additionally, the air temperature anomalies associated with the leading sea ice EOF shown in Fig. 9 are similar to those associated with the NAO–AO (Fig. 5) although they have a stronger signal over the Greenland and Barents Seas.

Air temperature, surface wind, and presumably ocean circulation changes are associated with both the observed and modeled sea ice variability, and indicate that both dynamic and thermodynamic mechanisms are likely involved in the forcing of the ice anomalies. The effects of these mechanisms can be diagnosed directly from the simulation results, whereas the equivalent observational analysis is limited by inadequate data. More specifically, during run time, model diagnostics of the grid area tendency due to dynamic and thermodynamic processes are computed and saved. These wintertime ice area tendency terms are averaged for Labrador Sea, Barents Sea, and Greenland Sea region. These regions coincide with areas where the anomalies associated with the first sea ice EOF (Fig. 7a) exceed 10% for the Labrador Sea and Barents Sea and exceed 5% for the Greenland Sea. The thermodynamic pro-
cesses responsible for changes in ice area include all of the melt/growth terms including basal, lateral, and surface processes. The dynamic processes are everything else that acts to change the ice area, including advection of sea ice and changes in area due to ridging and rafting. Table 1 shows the regression of the ice area tendency terms on the summer ice index, defined as the principal component of the leading mode of JFM sea ice variability, for the preceding autumn (OND) and coincident winter (JFM) seasons. In the Labrador Sea, where enhanced ice cover is associated with the NAO–AO, both dynamic and thermodynamic processes force the ice area changes in autumn, with the thermodynamic processes being considerably more important. This is consistent with ice–ocean mixed layer hindcast simulations that are forced with atmospheric reanalysis data (Deser et al. 2002). Similar mechanisms are at work in the Barents Sea, where reduced ice cover is associated with the NAO–AO. During the winter in both the Labrador and Barents Seas, the thermodynamic process of anomalous ice growth continues to increase the ice area anomalies whereas dynamic processes reduce them, acting as a negative feedback. In the Labrador Sea, these dynamic processes are manifested as enhanced southward ice advection out of the maximum ice anomaly region into relatively warm water where the ice consequently melts. In the Barents Sea, because less ice is present during a high NAO–AO index, less ice is available to be exported out of this region, resulting in the negative dynamic feedback. In the Greenland Sea, where negative ice area anomalies are associated with a positive NAO–AO, the ice concentration anomalies are thermodynamically driven and dynamical processes, essentially enhanced ice transport into the region, act to increase the ice cover and reduce the anomalies in both autumn and winter.

2) SUMMER VARIABILITY

There have been substantial and significant trends in observed summer ice conditions over the last several decades (e.g., Johannessen et al. 1995; Maslanik et al. 1996; Parkinson et al. 1999). The observed changes are largest in the East Siberian and Laptev Seas and appear to be related to an increased frequency of springtime cyclones over the central Arctic (Serreze et al. 1995; Maslanik et al. 1996). Deser et al. (2000) concur that the summer ice anomalies are related to springtime atmospheric conditions and show that these are related to the NAO–AO. However, the relationship between spring SLP anomalies over the Arctic and the wintertime NAO–AO index is not consistent over the observed record, indicating that other factors are contributing to the forcing of summertime ice anomalies. This agrees with other studies (e.g., Serreze et al. 1997; Rogers and Moseley-Thompson 1995) that suggest that the sea level pressure changes over the high Arctic are influenced by mechanisms other than the NAO–AO.

Figure 10a shows the first EOF of sea ice concentration for summer (JAS), which explains 20% of the variance. The pattern has some similarities to the recent trends in ice concentration, with maximum anomalies in the Siberian and Laptev Seas and shows considerable similarity to the observed EOF (Fig. 10b). However, the observed JAS ice variability exhibits reductions that are more closely confined to the coast and are also present in the Beaufort Sea. Additionally, the observed positive ice anomalies show a stronger signature in the Barents Sea than is present in the model simulations. Some of these discrepancies are probably related to biases in the simulated mean state of the ice cover. The simulated climatological ice cover is relatively thin, which may allow the anomalies to extend further into the basin. Additionally, the simulations have reduced ice cover in the Barents Sea compared to observations, leading to reduced variability in this region.

The simulated summer ice anomalies appear to be driven by springtime (AMJ) SLP conditions. The springtime SLP pattern associated with the leading mode of simulated ice variability is shown in Fig. 11a. It indicates that there is a decrease in SLP along the Eurasian coast and a small increase along the North American coastline. This is consistent with anomalous ice transport from the Siberian/Laptev Seas toward Fram Strait and the Barents Sea. This analysis compares very well to the equivalent analysis from the observations (Fig. 11b), indicating that realistic mechanisms are forcing the simulated variability. Additionally, this analysis of the simulations is consistent with the observational studies of Rigor et al. (2002) and Deser et al. (2000).

The relationship between this leading mode of summertime ice variability and the simulated winter NAO–AO index is minimal. They are only correlated at 0.14. In order to more directly examine the influence of the wintertime NAO–AO on the simulated summertime ice concentration a regression analysis is performed. Figure 12 shows the summer (JAS) ice con-

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**Table 1. Regression of the ice area tendency terms on the JFM ice index for different seasons and regions.** Here Th refers to thermodynamically forced area tendency, Dyn refers to dynamically forced area tendency, and Total refers to the total tendency. The values are in km$^2$ day$^{-1}$ (std dev)$^{-1}$ of the ice index.

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<tr>
<th>Region</th>
<th>Th</th>
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<th>Total</th>
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<tbody>
<tr>
<td>Labrador Sea</td>
<td>OND</td>
<td>40 408</td>
<td>29 374</td>
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<td></td>
<td>JFM</td>
<td>88 294</td>
<td>-44 412</td>
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<td>Greenland Sea</td>
<td>OND</td>
<td>-11 563</td>
<td>37 22</td>
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<td></td>
<td>JFM</td>
<td>16 805</td>
<td>12 209</td>
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<tr>
<td>Barents Sea</td>
<td>OND</td>
<td>-21 835</td>
<td>-14 194</td>
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<td></td>
<td>JFM</td>
<td>-72 205</td>
<td>58 811</td>
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centration regressed on the simulated winter NAO–AO index. It indicates that there is a decrease in ice cover in the East Siberian Sea during a high NAO–AO index. These anomalies are led by relatively thin winter ice cover. The winter ice volume rate of change associated with one NAO–AO standard deviation is $-2.0 \text{ km}^3 \text{ day}^{-1}$ for dynamically forced changes and $0.6 \text{ km}^3 \text{ day}^{-1}$ for thermodynamically forced changes. This suggests that the NAO–AO associated ice volume anomalies within the East Siberian Sea are driven by anomalous ice advection associated with the NAO–AO wind forcing. In fact, there are considerable changes in ice motion associated with the NAO–AO (Fig. 13). The simulated ice velocity changes are largest in winter (Fig. 13a) but continue into the summer (Fig. 13b). The spatial patterns

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**Fig. 10.** The first EOF of (a) simulated and (b) observed summer (JAS) sea ice concentration. The contour interval is 5% and the zero contour is not shown. Negative values are shaded. The observational analysis uses SSM/I satellite passive microwave data (Cavalieri et al. 1999) from 1979 to 1998.

**Fig. 11.** The spring (AMJ) sea level pressure regressed on the JAS ice concentration EOF for (a) the simulation and (b) the observations. The contour interval is 0.5 mb, negative values are dashed, and the zero contour is not shown.
and magnitudes of these velocity anomalies generally agrees with the observational analysis discussed by Rigor et al. (2002). However, the changes in summer velocity associated with the NAO–AO are displaced compared to the observations, with the observations showing that the center of the anomalous gyre occurs in the vicinity of the North Pole. This displacement may lead to discrepancies in the simulated and observed relationship between the NAO–AO and the summer ice conditions. In particular, it appears that the simulated relationship between summer ice variability and the NAO–AO is weaker than indicated by the observations.

The anomalous ice advection associated with the NAO–AO in the model solutions is somewhat different to that associated with the first EOF of summer ice concentration (Fig. 10a). In particular, there is reduced transport of ice from the Beaufort Sea to the East Siberian Sea associated with the NAO–AO in contrast to an increased transport of ice away from the East Siberian–Laptev Seas associated with the summer ice EOF. This results in ice concentration anomalies in Fig. 12 that are considerably more coastally confined than those shown in Fig. 10a. The changes in ice motion associated with the NAO–AO or the summer ice EOF typically lead the summer ice variability and cause changes in winter and spring ice volume. Thermodynamic processes damp the spring ice volume anomalies, because the presence of thinner and less concentrated ice cover increases the ice growth rates. However, the dynamic processes dominate and because thin ice cover is more easily removed during the melt season enhanced summertime open water formation follows the spring ice volume anomalies. This enhanced summertime ice

Fig. 12. The simulated summer (JAS) ice concentration regressed on the NAO–AO index. The contour interval is 2% and the zero contour is not shown. Negative values are dashed.

Fig. 13. The simulated (a) winter (JFM) and (b) summer (JAS) ice velocity regressed on the NAO–AO index. Note that the references vector, shown below the individual panels, is different for the winter and summer regressions.

JFM Regression

(a) Ice velocity cm/s/σ

180
150W
150E

120W
120E

90W
90E

60W
60E

30W
30E

(b) JAS Regression

Ice velocity cm/s/σ

180
150W
150E

120W
120E

90W
90E

60W
60E

30W
30E

Fig. 13. The simulated (a) winter (JFM) and (b) summer (JAS) ice velocity regressed on the NAO–AO index. Note that the references vector, shown below the individual panels, is different for the winter and summer regressions.

Thus, while the NAO–AO is not responsible for a majority of the variability in the simulated summer ice concentration in the model, it does contribute to changes in ice area within the Siberian Sea. This is largely driven by winter ice volume changes that result in enhanced
3) Ice Export Variability

Kwok and Rothrock (1999) show that ice export through Fram Strait from 1978–96 can be directly linked to the NAO index with larger ice export corresponding to a high NAO index. However, in a hindcast modeling study that examined ice export variations from 1958 to 1998, Hilmer and Jung (1999) argued that there was essentially no correlation to the NAO from 1958 to 1977. An analysis of the SLP from this earlier time period reveals that the anomalies associated with the NAO had a different spatial structure, with the low pressure center located southwestward of the more recent location and aligned more meridionally. This resulted in a negligible influence of the NAO on pressure gradients across the Fram Strait and hence the wind forcing in this region.

The mean value of the annually averaged simulated Arctic ice export is 0.07 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$), which is lower than the observed value of 0.09 Sv (Aagaard and Carmack 1989). This is due to the relatively thin Arctic ice cover that is present in the simulations (BRI). The standard deviation of the time series is 0.016 Sv. Changes in ice export are associated with variations in both the velocity and thickness of the ice transported through Fram Strait. As such, it is expected that variations in SLP and wind forcing would be associated with the ice export variability. The SLP regressed on the ice export time series is shown in Fig. 14. It exhibits reduced SLP along the Eurasian coast and increased SLP gradients across Fram Strait, leading to higher wind forced Fram Strait ice transport.

While somewhat higher ice velocities within the Fram Strait region are associated with the simulated NAO–AO index (Fig. 13), as indicated in Fig. 2, the NAO–AO has a relatively small influence on the SLP gradient across Fram Strait. This, in addition to a weak influence of the NAO–AO on ice thickness in the Fram Strait region, results in a low correlation between the Arctic ice volume export and simulated NAO–AO index of 0.15. As occurs in the observations, there are subsets of the full time series for which considerably higher correlations are present. For example, Fig. 15 shows the NAO–AO SLP pattern for a 50-yr subperiod of the integration (years 471 to 521). During this time the NAO–AO index has more influence on the pressure gradients across the Fram Strait and is correlated to the ice export there at 0.46 (which is significant at the 99% level). However, it is possible that this relatively high correlation is what would be expected purely by chance, and as this high level of correlation occurs rather infrequently for 50-yr segments of the entire integration length, it does not signify a robust relationship between the Fram Strait ice export and the NAO–AO. Thus, although large correlations between the NAO–AO and the Fram Strait ice export are present for subperiods of the integration and these appear to coincide with chang-
es in the spatial structure of the NAO–AO that project onto the SLP gradients across Fram Strait, we concur with the findings of Jung and Hilmer (2001) that there is only a weak link between the Fram Strait ice export and the NAO–AO variability.

4. Discussion and conclusions

The NAO, and the closely related AO, are the dominant modes of atmospheric variability in the North Atlantic and Northern Hemisphere, respectively. Observations suggest that the NAO–AO modifies Arctic sea ice and ocean conditions. The CCSM2 control integration has a realistic depiction of the pattern of the SLP anomalies associated with the NAO–AO. The primary discrepancy with observations is that the teleconnection to the North Pacific is considerably stronger in the model simulation. The simulated surface air temperature anomalies and precipitation anomalies associated with the NAO–AO are also very realistic. Associated with a positive NAO–AO index, the model simulates warm conditions across northern Eurasia and cold conditions centered over the Labrador Sea.

When considering separate 50-yr subperiods of the CCSM2 control integration, the centers of action associated with the simulated NAO–AO vary spatially. In particular, the low pressure center that usually occurs over southern Greenland, can also occur over the Barents and Kara Seas. This is consistent with observations that show a shift in the SLP pattern associated with the NAO over the last 40 yr. It is possible that the 50-yr subperiods from the model sample too few year-to-year variations to be considered robust. However, as the different spatial patterns often project onto relatively small “exchange” regions, such as the Fram Strait, considering these different time periods can lead to different conclusions on the influence of the NAO–AO on Arctic climate. The changes in the spatial pattern of the NAO–AO occur here in unforced climate integrations, suggesting that changing greenhouse gases are not necessary to drive such variability.

Although the temporal variability of the NAO–AO is reasonably simulated compared to the observed time series from 1864 to 1963, the recent observed 40-yr trend in the NAO–AO is never realized in the natural variability of the model solutions. If one assumes that the magnitude of the model variability is realistic, this implies that the increase in the observed NAO–AO is forced by external factors, the most likely factor being anthropogenic forcing. Future work will examine “climate of the twentieth century” simulations to see whether an increasing NAO–AO index such as that present in the observed record is realized when observed anthropogenic forcing is applied. However, it is also possible that processes not represented in the CCSM2, such as a well resolved stratosphere (e.g., Shindell et al. 1999), may be necessary to accurately simulate the recent observed NAO–AO trend.

The simulated winter sea ice variability compares well to observations. In particular, a dipole pattern with opposing centers of action in the Labrador Sea and Greenland–Barents Seas is present in the winter ice concentration. As in the observations, this variability is correlated to the NAO–AO with a positive index associated with extensive Labrador Sea and reduced Greenland and Barents Sea ice cover. The anomalies in the Barents and Labrador Seas are both dynamically and thermodynamically forced, although thermodynamic forcing plays the more important role. The Greenland Sea anomalies are thermodynamically forced with dynamic mechanisms acting to reduce the anomalies.

The dominant mode of summer ice concentration variability shows a center of action located in the Siberian sector of the Arctic with an opposing center located near Fram Strait. In agreement with observations (Deser et al. 2000), this variability is forced by springtime SLP anomalies that modify the ice transport. There is only a weak correlation of this variability to the NAO–AO over the entire time series. However, a regression of the NAO–AO onto summer ice conditions does indicate that it influences the ice cover in the Siberian Sea. In particular, during positive NAO–AO conditions, anomalous wind forcing reduces the transport of ice into the Siberian Sea. This results in thin ice cover in the region, which is readily melted the following summer, resulting in enhanced open water formation. This is consistent with observational (Rigor et al. 2002), hindcast modeling (Zhang et al. 2000), and idealized forcing modeling (Hu et al. 2002) studies. However, the simulations presented here suggest that while the observed increase in the NAO–AO contributes to the recent decrease in Siberian Sea ice cover, other factors are also important.

The ice export through Fram Strait is also influenced by the simulated NAO–AO. Although, in general, this influence is weak and is not consistent over the simulated record. High correlations are present for some subperiods, which are related to the changing spatial structure of the NAO–AO and how this influences pressure gradients across Fram Strait. However, this does not indicate a robust relationship between the NAO–AO and Fram Strait ice transport. These results agree with the hindcast modeling studies of Hilmer and Jung (1999) and Jung and Hilmer (2001).

Much of the work presented here has documented the simulation NAO–AO in the CCSM2 and its related high-latitude variability with comparisons to known observational relationships. This study has also provided new insight into mechanisms that influence Arctic sea ice variability and the robustness of some relationships in the context of the model variability. In particular, the interplay of dynamic and thermodynamic processes in modifying sea ice area and its response to the NAO–AO has been addressed. Additionally, as a long-simulated time series is available for this study, we have been able to examine spatial variations of the NAO–AO. Because small changes in the NAO–AO spatial
structure influence narrow exchange regions (such as the Fram Strait), examining different relatively short (e.g., 50 yr) time periods can lead to different conclusions on the influence of the NAO–AO on such things as the Fram Strait ice export. This points out the need for longer observational records of high-latitude climate conditions.

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