Atmospheric circulation and its effect on Arctic sea ice in CCSM3 simulations at medium and high resolution

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Abstract

The simulation of Arctic sea ice and surface winds changes significantly when CCSM3 resolution is increased from T42 (a 64x128 global grid) to T85 (a 128x256 grid). At T42 resolution, Arctic sea ice is too thick off the Siberian coast and too thin along the Canadian coast. Both of these biases are reduced at T85 resolution. The most prominent surface wind difference is the erroneous North Polar summer anticyclone, present at T42 but absent at T85.

An offline sea ice model is used to study the effect of the surface winds on sea ice thickness. In this model, the surface wind stress is prescribed alternately from reanalysis and the T42 and T85 simulations, and all other forcing inputs are calculated from fixed observational data. In the offline model, CCSM3 surface wind biases have a dramatic effect on sea ice distribution: with reanalysis surface winds annual-mean ice thickness is greatest along the Canadian coast, but with CCSM3 winds thickness is greater on the Siberian side. A significant difference between the two CCSM3-forced offline simulations is the thickness of the ice along the Canadian archipelago, where the T85 winds produce thicker ice than their T42 counterparts. Seasonal forcing experiments, in which CCSM3 (reanalysis) winds are used in the spring and summer (fall and winter), relate the Canadian thickness difference to surface wind differences in spring and summer. These experiments also show that the ice build up on the Siberian coast at both resolutions is related to the fall and winter surface winds.

The Arctic atmospheric circulation is examined further through comparisons of the winter sea level pressure (SLP) and eddy geopotential height. At both resolutions the simulated Beaufort high is quite weak, weaker at higher resolution. Eddy heights show that the wintertime Beaufort high in reanalysis has a barotropic vertical structure. In contrast, high CCSM3 SLP in Arctic winter is found in association with cold lower-tropospheric temperatures and a baroclinic vertical structure.

In reanalysis, the summertime Arctic surface circulation is dominated by a polar cyclone, which is accompanied by surface inflow and a deep Ferrel cell north of the traditional polar cell. The Arctic Ferrel cell is accompanied by a northward flux of zonal momentum and a polar lobe of the zonal-mean jet. These features do not appear in the CCSM3 simulations at either resolution.
1. Introduction

With its triple role as shortwave optical reflector, air-sea flux barrier, and salt transporter, Arctic sea ice plays a key role in simulations of anthropogenic climate change. As a recent example, Holland and Bitz (2004) found that climate models with relatively thin sea ice showed more northern hemisphere polar amplification of global warming than those with thicker ice. This finding is in keeping with earlier studies (e.g. references cited in Houghton et al. 2001) which show a prominent role for sea ice in setting the climate sensitivity of the Arctic and even the globe as a whole (Rind et al. 1995). But accurate sea ice simulation is hard to achieve, since ice evolves through complex interactions of mixed phase saline thermodynamics, radiative transfer, and rheology. Moreover, even a perfect sea ice model will generate errors given a poor simulation of the overlying atmosphere. In particular, the spatial pattern of sea ice thickness is largely determined by surface winds.

While possibly less mysterious than the sea ice, the Arctic surface winds are also poorly understood, and current climate models have difficulty in reliably simulating them. In a study of several simulations from the Atmospheric Model Intercomparison Project (AMIP; Gates 1992), Bitz et al. (2002) found a high bias in wintertime sea level pressure (SLP) off the central Siberian coast, with an accompanying anticyclonic bias in geostrophic surface winds. In an offline sea ice model, they found that the wind bias produced a sea ice pattern with maximum thickness along the Siberian coast, diametrically opposed to the Canadian-side pile up expected from observations. Weatherly et al. (1998) found a high bias in annual-mean SLP over the Beaufort Sea in the NCAR Climate System Model (CSM), a precursor to CCSM3 with more simplistic sea ice dynamics. The associated surface winds forced a sea ice pattern with maximum thickness against the Bering Strait, a 90° rotation from the expected Canadian thickness maximum. A study of an earlier generation of climate models by Walsh and Crane (1992) also found substantial errors in Arctic SLP, with seasonal pattern correlation coefficients between observed and simulated climatological SLP ranging from 0.9 to 0.17.

Previous studies have argued that wind-induced thickness errors derive from errors in the position and strength of the wintertime (DJF) Beaufort high (Bitz et al., Weatherly et al.). Yet models also have
severe errors in summer surface circulation (Bitz et al. Briegleb and Bromwich, 1998) which could be quite detrimental to sea ice thickness. Some support for a summer influence can be inferred from analysis of observations by Rigor et al. (2002), who found that decadal SLP changes in the summer and winter seasons were accompanied by comparable changes in sea ice motion. They noted that while the summer SLP change is smaller, the internal stress of the ice is also less for the thinner summer ice, so that the ice becomes more sensitive to wind stress as the surface winds slacken. The enhanced sensitivity of the summertime sea ice was also invoked by Serreze et al. (1989), who argued for a 20% reduction in ice concentration in the Canada Basin following extended periods of cyclonic summertime surface winds (see also Hibler 1974). The summer sensitivity was quantified by Thorndyke and Colony (1982), who found that, for geostrophic surface winds of the same strength, the response in sea ice motion was stronger in summer than in winter by a ratio of 11 to 8.

Our interest in the seasonality of the surface wind errors comes in large part from our desire to link Arctic surface wind forcing of sea ice to the large-scale general circulation. The wintertime Beaufort high can be regarded as a stationary wave, clearly apparent in plots of, say, the eddy geopotential height at 1000 and 700 mb. (section 5). Attempts to account for the climatological stationary waves usually involve consideration of the strength of the zonal-mean flow, diabatic heating in stormtracks and in the tropics, vorticity flux by transients, and stationary nonlinear interactions. Held et al. (2002) present evidence that the wintertime stationary waves in high latitudes – including the Beaufort high – depend on the interaction of tropical heating with midlatitude mountains (their figure 11). A better understanding of these dynamics could be helpful in understanding why the Beaufort high is severely distorted in GCMs. In contrast, the observed summer circulation of the Arctic is much more zonally symmetric, dominated by a polar low and flanked by the stormtrack of the Arctic front (e.g. Serreze et al. 2001; Serreze and Barry 1988; Reed and Kunkel 1960). The ability of a model to correctly simulate the wintertime Arctic circulation features may thus be quite independent of its ability to simulate the summertime flow.

Motivated by the difficulty and importance of accurate sea ice simulation, the present study has three goals: 1) to identify the biases in Arctic basin surface winds and sea ice thickness in the NCAR CCSM3; 2)
to examine in detail the seasonally varying forcing of the sea ice thickness distribution by the surface winds; and 3) to relate the basin-scale surface wind biases to the large-scale CCSM3 atmospheric circulation. Our examination considers sea ice and circulation differences between two control run integrations of CCSM3, one at relatively high T85 resolution and one at a medium T42 resolution (see section 2 for specifications).

Although neither model produces an entirely satisfactory simulation of Arctic sea ice thickness, some improvement is evident at the higher resolution, as discussed in section 3. One might naively expect that more resolution is always better, since smaller grid spacing is presumably a better representation of the continuous atmosphere (and the detailed topography beneath it). This notion is especially tempting in polar regions, since resolution near the poles is compromised in spectral models like CCSM3. On the other hand, Bitz et al. did not find that higher resolution consistently yielded better Arctic SLP simulations among the AMIP integrations they examined. Furthermore, the physical parameterizations of CCSM3 are retuned at each resolution, so improvements may not be entirely due to the increased resolution.

Are improvements in sea ice thickness due to genuine improvements in surface winds? If so, what are the most relevant improvements? We examine the surface winds at the two resolutions to address these questions. We also use an offline sea ice model to isolate the mechanical effect of surface winds on the sea ice distribution. In the offline simulations, the surface stress on the top of the ice is calculated either from reanalysis or CCSM3 surface winds, but all other forcings (e.g. air-ice fluxes of latent and sensible heat) are computed from fixed observational datasets (details in section 2.3).

The remainder of this paper is divided into 6 sections. Section 2 discusses the model integrations and reanalysis datasets used in the study, and describes the offline sea ice model which we use to assess the effects of spring and summer surface wind biases on sea ice thickness. Section 3 compares the sea ice and surface winds between reanalysis and the CCSM3 integrations, and section 4 presents the offline sea ice model experiments used to relate the winds to the ice thickness. Section 5 documents the three dimensional structure of the observed and simulated DJF Beaufort high. This analysis is intended as a first step in understanding how the Beaufort high at the surface is linked to the large-scale circulation of the Northern Hemisphere winter. In section 6, we consider the summertime circulation, and show that, despite
the improvement at T85, neither resolution captures the North Polar summer low. Conclusions follow in section 7.

2. Data sources and model integrations

2.1. Reanalysis data

Reanalysis data for this study come primarily from the National Centers for Environmental Prediction (NCEP) – National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al., 1996). Monthly-mean wind, height, and temperature data were obtained from the NCEP website (currently nomad2.ncep.noaa.gov /ncep_data) on 17 pressure levels, and reduced to a 5° longitude by 2.5° latitude grid. The actual resolution of the height, wind, temperature, and vertical velocity fields is somewhat less than the grid spacing, since NCEP truncates these fields spectrally during postprocessing at total wavenumber 36 (T36; truncation discussed below). In addition to the monthly-mean fields, the meridional flux of zonal momentum by submonthly transients was obtained from the same site. For the discussion of low- and bandpass- frequency contributions to the northward momentum flux, daily-mean zonal and meridional winds at 200mb were downloaded from www.cdc.noaa.gov /cdc/reanalysis/reanalysis.shtml. Daily means were used for the flux calculations instead of instantaneous values to facilitate comparison with the archived daily data available for the model integrations. Monthly-mean surface winds were also obtained from this website.

Reanalysis data were also obtained from the European Center for Medium-Range Weather Forecasting (ECMWF 1997) 40-year reanalysis, referred to here as ERA-40. For most fields (e.g. sea level pressure and zonal winds) the two reanalyses agree quite closely, the exception being the vertical motion fields in figure 8. ERA-40 data shown here is archived on a 2.5° × 2.5° latitude-longitude grid.

The period of record for the reanalysis statistics shown here is 1980 to 1999. For consistency, we used 1979 to 1998 for the December months, to keep contiguous DJF seasons. The period was chosen to avoid possible differences in meteorological data relating to the advent of the satellite observations. Also, Rigor et al. (2002) note significant changes in Arctic SLP in both summer and winter between the periods 1979 to 1988 and 1989 to 1998, largely related to the Arctic Oscillation (Thompson and Wallace 2000; Thompson et
By averaging over the 20-year period, we avoid the somewhat arbitrary decision of which sub-period should be compared against the climate model.

2.2. CCSM3 integrations

Climate model data used here comes from the NCAR’s Community Climate System Model version 3 (CCSM3), which consists of atmosphere, land surface, ocean, and sea ice components which communicate with each other through a flux coupler. Detailed descriptions of the component models and the flux coupler can be found in references in this issue, as well as overviews of the atmospheric and oceanic circulations and global climate (Collins et al. 2005a,b; Dickinson et al. 2005, Holland et al. 2005, Large et al. 2005).

We examine data from model integrations B30.009 and B30.004, control runs for the anthropogenic global warming scenarios specified by the Intergovernmental Panel on Climate Change (IPCC). In the control runs, radiative forcings (e.g. concentration of carbon dioxide) are held fixed at 1990 levels during a 1000-year integration of the model. The primary difference between the two control runs is a factor of two (in each direction) increase in the horizontal resolution of the component atmosphere model, referred to as the Community Atmosphere Model version 3 (CAM3), in the B30.009 integration. CAM3 is a spectral model with triangular truncation, in which resolution is expressed as the total wavenumber of the smallest-scale spherical harmonic function used to represent the atmospheric state variables. B30.009 is a T85 integration, meaning that spherical harmonic functions with total wavenumber less than or equal to 85 are used to represent the atmospheric state, while B30.004 is a T42 integration. The T85 resolution uses a 128 × 256 latitude - longitude grid, while T42 uses a 64 × 128 grid. Both resolutions use 26 levels in the vertical, and the resolution of the ocean, land, and sea ice models is the same for both integrations at 1.125° in longitude and ~0.5° in latitude, except in the tropics where the latitudinal resolution is finer. In addition to the resolution difference, the integrations differ in the tuning of various physical parameterizations. The precise details of this tuning are not easily ascertained.

Climatologies of sea ice and atmospheric circulation fields shown here were calculated using years 200 to 219 of the model integrations. The period of record is somewhat arbitrary, but it occurs after the large spin-up of Northern Hemisphere ice volume during years zero to 100, as described in Collins et al. (this issue).
While the period is somewhat short, the key features of the sea ice and circulation are consistent with those of other periods. For example, sea ice thickness plots for years 401 to 410 available at www.cccsm.ucar.edu/experiments/ccsm3.0 show a very similar difference pattern between the two integrations.

2.3. The offline sea ice model

The offline sea ice model is an updated version of the one used in Bitz et al. (2002). Specifically the model now includes a subgrid-scale parameterization of the probability density function of ice thickness, known as an ice-thickness distribution model, using the method of Bitz et al. (2001). The physics in the offline model is essentially identical to that in the sea ice component of CCSM3, which uses the same ice-thickness distribution method. However the offline model uses a first order accurate numerical solutions for horizontal advection with the ice motion field and for the so-called thickness advection of the ice-thickness distribution that results from ice accretion or ablation. Twice as many thickness categories are used in the offline model to compensate for the reduced accuracy of the numerics. The offline model grid is 80km square Cartesian.

3. Sea ice and surface winds

3.1. Sea Ice Thickness

Figure 1 shows the sea ice thickness climatology for CCSM3 at T85 (left) and T42 (right) resolutions, contoured at 0.5m intervals at all points where ice covers at least 3% of the grid box (in the CCSM3 output ice thickness is defined as ice volume divided by the total area of the grid box). The annual-mean thickness patterns (top row) show that the ice is generally thicker at the lower resolution, which has a local sea ice maximum in excess of 3m in the central Arctic, where the T85 sea ice is always less than 2.5m. Substantial differences in the pattern of ice thickness are also apparent. On the Siberian side, both resolutions have a local thickness maximum on the eastern coastline near Wrangell Island (near the dateline), which is much thicker and more extensive in the T42 integration. There is also a substantial difference in the onshore thickness gradient at the Canadian coast, where ice thickness increases from less than 2.5m to values locally in excess of 3.5m in the T85 integration. In the T42 case the thickness distribution is flat near the coast, with
no Canadian pile up. Ice builds up at the northern tip of Greenland in both cases, with a greater buildup at the higher resolution.

To examine the seasonal cycle of the ice thickness, we display in the bottom two rows the thickness patterns for April (middle row), and September (bottom row), the extreme months of the cycle. The patterns at both extremes are similar to the annual mean, with the same preference for more ice on the Canadian coast and less on the eastern Siberian coast at higher resolution. In September the T42 run has a narrow strip of thin ice (less than 2m thick) along the coast of the Canadian Archipelago (bottom left), while the T85 run has a thickness maximum at the same location (bottom right).

3.2. Surface winds

In figures 2 and 3 we examine the surface winds over the Arctic basin for the T85 and T42 simulations and the NCEP/NCAR reanalysis, with the expectation that differences in surface winds can be related to differences in ice thickness. It must, of course, be noted that the ice motion will be slightly to the right of the surface wind (e.g. Hibler and Flato 1992 fig. 12; Thorndyke and Colony 1982; Serreze et al. 1989, all of whom calculate turning angles from the geostrophic wind rather than the actual wind). Figure 2 shows surface wind vectors for the four seasons (DJF, top row; MAM, second row; JJA, third row; and SON, bottom row) at T85 (left column) and T42 (middle column). The difference T85 minus T42 is also shown in the right column. These are winds from the lowest model level, at which ambient pressure divided by surface pressure is 0.992 (the model uses a hybrid vertical coordinate which reduces to a standard sigma coordinate at the surface). In all panels the wind vector is suppressed at the pole to prevent plotting problems related to the convergence of the meridians.

Surface winds at the two resolutions have much in common, including northerly flow from the Arctic into the North Atlantic in all seasons, an anticyclonic circulation extending across the basin from the Siberian coast in DJF, and flow from the Canadian Archipelago to eastern Siberia in SON. However, there are also significant differences between the two model runs, including differences in the anticyclonic circulation extending outward from the Siberian coast, which is stronger in DJF in the T85 integration but stronger in SON in the T42 simulation. Also, the winds blowing from the Arctic into the North Atlantic are stronger at
T42 in all seasons except DJF, a difference which could be consequential for the export of sea ice from the Arctic. A pronounced difference between the runs is the JJA polar anticyclone, which is present in the T42 run but absent at T85. The wind difference plots show that in all seasons except DJF the T85 winds have a component toward the Canadian islands relative to the T42 winds. This wind difference is consistent with the thickness patterns in figure 1, which show more ice on the Canadian coast for the higher resolution.

Figure 3 compares the T85 surface winds with surface winds from the NCEP/NCAR reanalysis. Reanalysis winds are from the lowest level of the model used to produce the reanalysis, where pressure divided by surface pressure is 0.995. NCEP/NCAR winds are shown for each season in the right column, with the wind difference T85 minus reanalysis in the left column. It is clear from the plot that wind differences between T85 and reanalysis are as strong as the winds themselves. Furthermore, the T85 winds are generally directed more away from the Canadian Arctic and toward eastern Siberia than the reanalysis winds. Comparing this bias with the wind difference in figure 2, one can see that the T85 run amounts to a partial correction of the bias in the T42 run, in the sense that the T85 winds are directed more toward the Canadian side than the T42 winds, but still less toward Canada than the reanalysis winds. In the summer circulation, reanalysis winds have a cyclonic circulation around the pole. This is exactly the opposite of the anticyclone in the T42 integration. Thus we can say that that even though the T85 run does not produce the erroneous anticyclone found at T42, it still does not capture the polar cyclone found in reanalysis.

4. Offline sea ice model experiments

To quantify the effect of the surface wind on the ice we use an offline sea ice model, which uses the same ice rheology and thermodynamics as CCSM3 but isolates the influence of the surface winds. Our modeling strategy is to first generate a thickness pattern using winds from the NCEP/NCAR reanalysis, then generate comparable patterns using CCSM3 surface winds at both resolutions. As discussed in section 2.3, observationally derived values are used for all inputs other than the surface wind stress, so that the differences between offline simulations can be ascribed entirely to wind forcing. In addition, only the geostrophic component of the surface winds is used, so that simulation differences can be related to the sea
level pressure differences discussed in section 5.

4.1. Offline thickness comparisons

Figure 4 shows the thickness pattern generated by the offline model forced with NCEP/NCAR surface winds. The figure shows sea ice thickness for April (left panel) and September (middle), the extreme months of the annual cycle, and the annual mean (right). In all panels the thickest ice is along the Canadian coastline, with values in excess of 4m along the whole coast and local maxima in excess of 5m between Ellsmere Island and Greenland and off the islands of the Canadian Archipelago. Thickness decreases outward across the Arctic basin from the Canadian side, although a second local maximum is apparent in the East Siberian Sea. The annual cycle, assessed by comparison of the left and center panels, is most evident in the cross-basin thickness gradient, as ice thins from April to September along the Siberian and Alaskan coastlines but remains thick on the Canadian coast. The pattern of annual-mean sea ice thickness in the model is similar, although with reduced amplitude, to the pattern in Bourke and Garrett (1987; also shown as fig. 8b of Weatherly et al. 1998), who used submarine data to estimate the thickness pattern. The amplitude difference is partly a matter of definition, as the thickness shown here is ice volume divided by the area of the grid box, while Bourke and Garrett excluded ice-free areas from their mean thickness.

Comparison of the annual mean patterns in figures 1 and 4 shows a substantial discrepancy, as the CCSM3 sea ice does not show the strong preference for thicker Canadian-side ice produced by reanalysis winds. Of the two resolutions, The T85 integration (fig. 1, top right) is a better match with the offline reanalysis pattern in that it produces a thickness maximum along the Canadian coast, the T42 mid-basin maximum is gone, and the Siberian pile up is much reduced. On the other hand, the higher resolution still fails to produce the pronounced cross-basin thickness gradient found in the offline calculation.

Thickness patterns generated by the offline model forced with surface winds from CCSM3 at T85 (top row) and T42 (bottom row) resolution are shown in figure 5. At both resolutions, the annual-mean thickness patterns (right column) are diametrically opposed to the reanalysis pattern (fig. 4 right), with thickness increasing across the middle of the basin from the Canadian to the Siberian Arctic. The comparison thus shows that even if all temperature biases in the model were eliminated, the surface wind biases would be
sufficient to drive a thickness distribution in general opposition to the pattern implied by reanalysis winds. This opposition can be understood in terms of the surface wind differences between reanalysis and CCSM3 in figure 3, since the wind difference between T85 and reanalysis is directed toward Siberia in all seasons except spring (MAM).

Intercomparison of the annual-mean thickness patterns for the two resolutions, in figure 5, reveals substantial differences between the pattern forced by the T42 (top) and T85 (bottom) geostrophic surface winds. Sea ice is thicker overall at the higher resolution, with values in excess of 3m across the basin, or about 0.25m thicker than at T42, in opposition to the mean thickness difference of CCSM3 sea ice (figure 1, top row). Differences in mean thickness are usually attributed to thermodynamics (e.g. Randall et al. 1998), so the fact that the 2 meter temperature is colder at T42 than at T85 by about 2K in all seasons except summer (when temperatures are near the melting point) should be important in promoting more ice growth at the lower resolution. For the offline calculations, in which atmospheric temperature is the same, the difference can be attributed to nonlinear feedbacks between ice growth and ice motion (e.g. Zhang et al. 2002). Sea ice export from the Arctic is one contributing factor: about 5% more ice (by volume) is exported through Fram Strait (between Greenland and Spitzbergen) in the offline T42 case due to stronger surface winds blowing from the Arctic to the North Atlantic (fig. 2). M. Holland (personal communication) calculated that the average CCSM3 Fram Strait ice export is in excess of 0.096 Sverdrups (Sv) at T42, while about 0.087 Sv were exported at T85, consistent with the offline export difference (her calculation was for years 450-499 of the integration).

Further inspection reveals that the two resolutions differ in the extent of the Siberian pile up, with ice thickness in excess of 3.5m along the Siberian coast between Wrangell Island and the New Siberia Islands (Novosibirskaya Ostrova, near 145°E) in the T85 run. In contrast, thickness forced by the T42 surface winds has a local maximum offshore, with ice about 3m thick along the coast. The offshore T42 thickness maximum is consistent with the online T42 thickness in figure 1. However, the finding of thicker ice along the coast at the higher resolution is in opposition to the CCSM3 sea ice output, which shows larger Siberian thickness at the lower resolution. It would be difficult to anticipate the relative extent of the Siberian pile up
from the surface winds in fig. 2. In MAM, JJA, and SON, the T85 winds appear to be more offshore at the eastern end of the Siberian coastline, but significant thickness discrepancies extend further westward, where the wind differences tend to be oriented along the coast. In DJF, the T85 surface winds are directed more toward land than the T42 winds along most of the relevant coastline, but this is the season in which the ice should be least responsive to wind stress. While we cannot offer an explanation of the difference in Siberian thickness between the online and offline sea ice, the seasonal forcing experiments of the next section show that SON and DJF winds are largely responsible for determining Siberian-side thickness.

Offline ice thickness differences on the Canadian side are qualitatively consistent with the online differences in figure 1, as the T85 surface winds produce thicker ice than their T42 counterparts. In the annual mean, ice thickness decreases westward along the Canadian coast, with values below 2m extending from the western archipelago to the Alaskan North Slope. In contrast, ice forced by T85 winds has thickness values in excess of 3m along the entire Canadian archipelago, with a small local maximum along the western side of the archipelago where ice produced by T42 winds is less than 2m thick. This secondary maximum is consistent with the ice pattern produced by reanalysis winds (fig. 4), although reanalysis winds produce a much stronger maximum. Differences along the Canadian coast are most pronounced in September, when a strong offshore thickness gradient occurs along the Canadian and Alaskan coastlines in the T42-forced simulation.

4.2. Seasonal forcing experiments

Any attempt to relate biases in ice thickness to biases in surface winds must take into account the strong seasonality present in both. In particular, the preference for thicker ice on the Siberian side in figure 5 is most evident in April (left column), following the winter season. Thus, although the CCSM3 winds are directed more strongly toward Siberia than the NCEP/NCAR winds in JJA, SON, and DJF, we speculate that the winds of the winter season are more important for establishing the excess Siberian thickness than the summer winds. On the Canadian side, the thinner ice at the T42 resolution is most evident in September (middle column), following the summer season. This suggests that differences on the Canadian side may be more strongly related to the spring and summer winds. The seasonality of the thickness biases in the offline
model is qualitatively similar to the thickness seasonality in CCSM3 output (fig. 1), so the offline model can provide insight into the seasonality of the wind-ice relationship in the climate model. Our interest in the seasonality of wind-forced thickness biases stems from our desire to relate these biases to the large-scale circulation, particularly the strength of the Beaufort high in winter and the spurious T42 polar anticyclone in summer.

In figure 6 we address the issue of seasonality by integrating the offline model using surface winds from reanalysis for the SON and DJF seasons and CCSM3 surface winds for MAM and JJA. Thus, the September, April, and annual-mean thickness plots in figure 6 differ from their counterparts in figure 4 to the extent that the spring and summer surface winds in CCSM3 differ from reanalysis winds in those seasons. Also, differences between figure 6 and figure 5 can be attributed to differences between CCSM3 and reanalysis winds in SON and DJF.

Comparison of the annual-mean patterns in figs. 4 – 6 show that the annual-mean thickness patterns produced by the seasonal forcing experiment more closely resemble the pattern produced purely by reanalysis winds than the patterns produced purely by CCSM3 winds. The closer resemblance to reanalysis thickness is mainly due to thinner ice on the Siberian side, which means that the cross-basin thickness gradient in figure 6 is in the same direction as the gradient in figure 4. Thus, the figure confirms our expectation that the preference for thicker ice on the Siberian side results primarily from biases in the fall and winter surface winds.

On the Canadian side, the use of fall and winter reanalysis winds does not qualitatively change the thickness differences between the two CCSM3 resolutions. As in figure 5, the T85 winds generate thicker ice on the Canadian side, while T42 winds produce relatively thin ice along the Canadian and Alaskan coastlines. This result is consistent with the seasonality of figure 5, in which the T42-forced ice thins dramatically near the coasts of Canada and Alaska. As a further check on the role of the JJA winds associated with the prominent erroneous anticyclone in the T42 case, we show in figure 7 the results of a seasonal forcing experiment in which reanalysis winds are used in all months except JJA. As in figure 6, ice thickness is reduced along the western Canadian and Alaskan coastlines in response to T42 winds, while the T85
winds produce a local thickness maximum on the west side of the Canadian Archipelago.

5. Structure of the Beaufort High

Attempts to find surface wind – sea ice relationships will inevitably hinge on the subtleties of small-scale wind features, like the the onshore component of the flow on opposite sides of the Arctic basin. On the other hand, the quality of atmospheric general circulation model (AGCM) simulations is typically judged by their depiction of large-scale circulation features, since the larger scales are most amenable to simulation by AGCMs. Are the relatively small-scale biases in the Arctic surface winds actually manifestations of deficiencies in the large-scale climatological circulation? In fall, winter, and spring, the surface winds of the Arctic basin are largely determined by the location and strength of the Beaufort high and the extension of the Icelandic low into the extreme North Atlantic, while in summer the surface winds circulate cyclonically around the polar low. Thus, we wish to relate the surface circulations in figures 2 and 3 to the simulations of these features.

Figure 8 compares the SLP in CCSM3 and the NCEP/NCAR reanalysis in all four seasons. The winter plots (left column) are dominated by the familiar Aleutian and Icelandic lows and the Siberian and Rocky Mountain highs, all of which are present in the CCSM3 integration, although the simulated oceanic lows are too low and the high over the Rockies is somewhat too strong. Over the Arctic, the DJF Beaufort high appears in the reanalysis (top row) as a ridge which connects the Siberian high with the high over the Rocky Mountains and separates the Aleutian and Icelandic lows. Such a ridge does not appear in the CCSM3 SLP at either resolution. Instead, the oceanic lows connect together across the Arctic basin, while the continental highs are separated by isobars which cross Alaska and eastern Siberia from the Arctic to the North Pacific. The SLP deficit associated with the missing Beaufort high is more pronounced at T85, since at that resolution SLP is lower over the Beaufort sea. The lower Arctic SLP at higher resolution is accompanied by deeper Aleutian and Icelandic lows than those found in the T42 integration. The deficiency of the simulated Beaufort high is also evident in MAM and SON, as the oceanic lows are too strongly connected in those seasons. The deficiency is most evident in MAM, when in reanalysis the Siberian high diminishes and
the Beaufort high emerges as a distinct, independent feature. As in the wintertime simulations, the Icelandic and Aleutian lows are deeper at T85 than at T42 resolution in MAM and SON.

The consequences of the SLP biases for surface winds can be appreciated by comparing figure 8 with figures 2 and 3. The wintertime Beaufort ridge in reanalysis is accompanied by geostrophic surface flow which crosses the Arctic basin from Siberia to Canada (fig. 3a). In the absence of the ridge, this cross-basin flow does not occur in the model, which instead produces a surface flow which leaves the Siberian coast near 90°E and returns to land near Wrangell Island on the eastern Siberian coast (fig. 2, top left and center). The surface wind bias toward the Siberian coast in the T85 simulation (fig. 3e), which plays a central role in generating the Siberian sea ice excess in figs. 1 and 5, can thus be interpreted as a consequence of the weakness of the Beaufort high. The same analysis applies at both resolutions for SON, and for the T85 integration in MAM.

The Beaufort high is a surface feature by definition, but we expect it to have a close association with the overlying atmosphere which could provide insights into the dynamics of the surface high. For example, the extreme cold surface temperatures of Siberia and central Asia are usually invoked in explanations of the Siberian high (e.g. Ramage 1971 section 3.1). This association is borne out by the baroclinic cold core vertical structure of the high. In addition, the anticyclonic surface winds circulating around the high must be maintained against friction by convergence of anticyclonic vorticity above the high, most of which occurs in the upper troposphere. As a tentative first step in understanding the dynamics of the surface high, we wish to identify the vertical structure of the flow associated with it, and determine whether the flow is best described as barotropic or baroclinic. Our variable of choice for this task is the eddy geopotential height (departure of geopotential height from its zonal average), plotted in conjunction with the eddy temperature field, from reanalysis data and CCSM3 output.

DJF Eddy geopotential height for the Northern Hemisphere extratropics is plotted in figure 9 for the NCEP/NCAR reanalysis at 1000mb (top left) and 700mb (top right), with a whitespace circle at 75°N (the reference latitude for panel c). The 1000mb plot (panel a) shows that the most prominent SLP features – the oceanic lows and continental highs – are well described by eddy height, as is the Beaufort high. These
features are also present at 700mb, but with the expected differences: the Aleutian and Icelandic lows are shifted westward, and the Siberian high is much reduced, consistent with its expected cold-core structure (e.g. Nigam and DeWeaver 2003). For our purposes, the most noteworthy feature is the Beaufort high, which appears at 700mb as a closed isobar, separated from the Siberian high but with some connection to the high over the Rockies. While the Siberian high diminishes from 1000 to 700mb and the Rocky Mountain high amplifies through the same layer, the Beaufort high has the same amplitude at both levels.

The vertical structure of the Beaufort high is further examined in figure 9c, a zonal-vertical cross section of the eddy height at 75°N in which the Beaufort high is found at and slightly to the east of the dateline. In addition to the height contours, the figure shows the eddy temperature field at that latitude. In agreement with the 1000 and 700mb plots the figure confirms the barotropic structure of the high, which has some amplification in the upper troposphere/lower stratosphere and an additional intensification in the boundary layer, where eddy temperatures are coldest. It is clear from the figure that, unlike the Siberian high, the Beaufort high is not closely associated with cold lower tropospheric temperatures, although the figure leaves open the possibility that the time-mean high can be viewed as a “graveyard of anticyclones” (e.g. Serreze and Barry 1988).

Eddy height from the CCSM3 integrations is shown in figure 10, with horizontal plots at 1000mb in panels a (for T85) and b (for T42). An advantage of eddy height as a marker for the surface highs and lows is that subtracting the zonal mean factors out the general tendency for lower pressure at the higher resolution. As with SLP in figure 8, comparison of eddy height in figures 9 and 10 shows that the Beaufort high is much weaker in the model than in the reanalysis. The vertical structure of the eddy height and temperature at 75°N is shown in panels c and d for the T85 and T42 resolutions, respectively. These panels show that the 1000mb highs at both resolutions are associated with the coldest lower tropospheric temperatures at that latitude, between 90°E and the dateline. In the T42 case the cold temperatures extend upward to 400mb and are associated with a strong upper-level trough which is much weaker in the higher resolution (in examining this figure it should be noted that $T \propto -\rho \partial \phi / \partial p$, so for a given temperature perturbation the vertical spacing of the height contours will be closer at lower pressure). The collocation of the upper trough and the surface
high demonstrate a clear cold-core structure which, though appropriate for the Siberian high, is not correct for the Beaufort high. There is, however, an upper-level ridge east of the dateline which is reminiscent of the upper part of the Beaufort high in figure 9c, so one could claim that the model is able to simulate the upper-level eddy height associated with the surface Beaufort high.

6. Maintenance of the zonal-mean summer circulation: the Arctic Ferrel cell

The summer circulation deserves separate treatment for two reasons. First, summer is the only season when the circulation cannot be easily described in terms of the Beaufort high and the oceanic lows. In the absence of strong zonal asymmetries, the structure and dynamics of the summertime flow can be conveniently portrayed through zonal averages of the relevant dynamical variables. Second, the summer surface circulation at T42 is in strong opposition to the reanalysis circulation, with a polar anticyclone instead of a cyclone. In this section we examine the underlying dynamics of the anticyclonic bias and the reasons for its reduction in the T85 integration.

Figure 11 shows the JJA zonal-mean SLP and surface zonal wind $\overline{\langle u_s \rangle}$, where the overbar and brackets represent averaging in time and longitude, respectively. The difference in JJA polar SLP between the T42 (represented dotted lines in all panels) and T85 (the dashed lines) integrations is evident in panel a, in which the T42 SLP has a maximum value of about 1016mb, comparable to midlatitude SLP values, while T85 SLP is close to 1010mb. Polar SLP from NCEP/NCAR reanalysis (panel b, solid line) is near 1009mb, so there is a clear improvement in SLP at the higher resolution. The dashed-dotted line in panel b represents zonal-mean SLP from the ERA-40 reanalysis, which agrees closely with the NCEP/NCAR profile. In both reanalyses, SLP has a qualitatively different structure than CCSM3, since reanalysis SLP has a local maximum at 70°N and a minimum at the pole.

The geostrophic consequence of the polar minimum is that reanalysis $\overline{\langle u_s \rangle}$ is westerly north of 70°N, with a maximum just above 1ms$^{-1}$, comparable to the strength of the midlatitude westerlies. In contrast, the T42 $\overline{\langle u_s \rangle}$ is easterly for all latitudes north of about 65°N, reminiscent of the polar easterlies of the standard three-cell general circulation model (e.g. Palmén and Newton 1969, figs. 1.2 and 4.1). T85 $\overline{\langle u_s \rangle}$ is a partial
correction of the T42 easterly bias, with weaker polar easterlies than T42 and weak westerlies between 80° and 85°N. A strong bias in CCSM3 $\vec{n}_s$ also appears in the midlatitudes, where at both resolutions the simulated surface westerlies reach a maximum of 3 ms$^{-1}$, about twice the maximum in reanalysis. Higher resolution does not significantly reduce this bias.

We next consider the mean meridional circulation ($[\vec{v}], [\vec{\omega}]$) accompanying the zonal-mean SLP profiles. Ekman balance requires that the cyclonic winds around a polar low have a poleward ageostrophic component, with rising motion over the pole to satisfy mass continuity. The overturning motion for a model with a polar SLP maximum would be in the opposite direction. These expectations are confirmed in figure 12, in which streamlines are used to show the sense of the Eulerian-mean meridional circulation in panels a – d, and $[\vec{\omega}]$ is shown in e – h. The streamlines were created using the GrADS plotting package, after converting $[\vec{\omega}]$ to z-coordinate vertical motion using $[\vec{\omega}] = -\frac{[\vec{u}]}{\rho} \frac{\partial [\vec{T}]}{\partial z}$, where $R$ is the gas constant, $[\vec{T}]$ is the JJA averaged temperature, and $g$ is the gravitational acceleration. We also multiply $[\vec{w}]$ by an arbitrary scaling factor of 350 to produce streamlines with comparable meridional and vertical spacing. Finally, The CCSM3 output includes pressure vertical velocity on the model vertical levels, which was linearly interpolated to pressure levels prior to calculating $[\vec{w}]$.

The sense of the meridional-vertical circulation for the two reanalyses (panels a and b) is similar, with the summer Hadley cell extending from about 15°N to the northern midlatitudes, and a Ferrel cell with rising motion around 60°N and subsidence below and to the south of the jet core (5, 15, and 25 ms$^{-1}$ isotachs of $[\vec{u}]$ are superimposed on the streamlines in a – d). Poleward of 60° the standard three-cell general circulation model has a single thermally direct polar cell, but here we find two cells: a thermally direct cell with subsidence near 75°N – the latitude of the SLP maximum in 11b – and a thermally indirect cell with rising motion at the pole. The strength and extent of the polar rising motion differs substantially between NCEP/NCAR and ERA-40, with much stronger NCEP/NCAR $[\vec{\omega}]$ values. Since the reanalysis SLP profiles agree quite closely (fig. 11b), differences in the strength of the mean meridional circulation could be due to differences in the effective surface friction. The approximate boundary layer momentum balance is between Coriolis force and friction, or $f[\vec{u}] - \tau[\vec{u}] = 0$, where $\tau$ is an effective Rayleigh friction
and $[\tilde{u}]$ is in geostrophic balance with the SLP gradient. If the SLP gradients and hence the $[\tilde{u}]$ values are the same, differences in $[\tilde{v}]$ would have to be balanced by differences in $r$. Alternatively, differences in mean meridional circulation given the same SLP gradient could be due to differences in mountain torque. In addition, differences in diabatic heating and eddy heat flux convergence are required to balance the adiabatic cooling differences implied by $[\tilde{w}]$ in panels a and b.

The CCSM3 mean meridional circulation (panels c and d) has Hadley and Ferrel cells in essentially the same locations as the reanalysis, but north of 60° we find a single thermally direct polar cell with subsidence at high latitudes. In panel g the T85 $[\tilde{w}]$ has a nodal line very near the pole, indicating that rising motion has been curtailed at the pole in association with the elimination of the polar anticyclone (fig. 11a). However, neither resolution produces the polar Ferrel cell found in reanalysis.

A further difference between reanalysis and CCSM3 zonal-mean circulations is the poleward lobe of the upper-tropospheric jet found above the latitudes of subpolar descent (around 75°N) in reanalysis. The existence of an Arctic westerly jet, distinct from the midlatitude jet, has been documented by Serreze et al. (2001), who showed cross-sections through the jet at several longitudes and associated it with the Arctic frontal zone. Of course, the dynamical relationship between the mean meridional circulation and the structure of the upper-level jets is somewhat indirect, and jet structure cannot be inferred from the mean meridional circulation or vice versa. Nevertheless it is noteworthy that without the polar Ferrel cell CCSM3 does not produce the poleward lobe at either resolution.

We have seen that in reanalysis the zonal-mean surface westerlies are almost as strong in the Arctic as they are in midlatitudes, and the T42 Arctic easterlies are comparable in strength to the reanalysis Arctic westerlies (fig. 11c). Since the surface stress associated with these westerlies constitutes a zonal momentum sink for the atmospheric column, convergence of zonal momentum flux is required to maintain the westerlies. Thus the surface wind differences in figure 11c suggest large differences in zonally averaged zonal momentum flux between reanalysis and the CCSM3 simulations.

These differences are examined in figure 13, starting in panel a with the zonally averaged meridional flux of zonal momentum by transients $[\tilde{u}'v']$, in which $u'$ and $v'$ are departures from the June-July-August
monthly-means of \( u \) and \( v \) averaged over the 20 years of record (for CCSM3 we used the average of the archived monthly-mean flux \( \overline{uv^{\text{m}}} \) for each month, minus the steady flux \( \overline{uv}^{\text{m}} \) for that month). The zonally averaged zonal momentum flux by the steady climatological \( u \) and \( v \), including zonal-mean and zonally asymmetric winds, is shown in panel b. In both panels, the mass-weighted vertical integral of the flux is calculated using pressure-level data between the surface pressure and 100mb, and the result is divided by an approximate mean column mass of 900kPa/\( g \) to produce a column-mean momentum flux with units of \( \text{m}^2\text{s}^{-2} \). In accord with the mean-meridional circulation in figure 12, the reanalysis transient momentum flux is poleward at high latitudes (from about 75° to 85°N), while the CCSM3 momentum flux is equatorward for all latitudes north of about 60°N. It is evident from panel a that, in addition to the opposite direction of the high latitude fluxes, the meridional flux of zonal momentum is considerably stronger in CCSM3 than in reanalysis. In the midlatitudes, just north of 40°, CCSM3 momentum flux is about 60% stronger, and in the region around 65°N where reanalysis and CCSM3 fluxes are both equatorward the CCSM3 flux is about twice as strong. CCSM3 momentum flux by the steady (panel b) flow is also stronger near 30°N, but in high latitudes the T85 steady momentum flux agrees closely with NCEP/NCAR flux. Finally, we note that while momentum flux from the T85 and T42 integrations are in close agreement in midlatitudes, there is a reduction of polar cap momentum flux in the T85 integration. This reduction is in general agreement with the reduction of the polar easterlies at T85.

Larger eddy momentum fluxes can be generated either by stronger eddy amplitudes or by a greater correlation between \( u' \) and \( v' \). In figure 13c we address the question of eddy amplitude by plotting the zonal average of the column-mean kinetic energy \([((u'^2 + v'^2)/2)]\) for the sub-monthly transients from panel a. According to panel c, midlatitude transient kinetic energy (TKE) is indeed stronger in CCSM3 than in reanalysis (and stronger at higher resolution), but the energy difference is smaller than the difference in momentum flux. For the T42 integration (the long-dashed curve) the midlatitude energy is about 15% higher than in the reanalysis, a small excess compared to the difference in momentum flux. North of 60°N the reanalysis TKE remains close to 60m^2s^{-2}, but the CCSM3 TKE decreases toward the pole. A strong high-latitude difference is evident between the two resolutions, since the T42 polar TKE is only half of
the reanalysis value (30m$^2$s$^{-2}$ versus 60m$^2$s$^{-2}$) while T85 TKE is roughly the same as in reanalysis. The difference in TKE is the opposite of the flux difference, since the lower resolution has stronger momentum flux than the higher resolution, in spite of the reduced TKE. Evidently the $(u',v')$ correlation (the meridional tilt of the eddies) is reduced as the eddies become stronger at higher resolution.

In panels d – f we further examine the eddies using bandpass and low-pass filtered winds from reanalysis and the T42 CCSM3 run at 200mb. 200mb was chosen as a representative upper-tropospheric level at which daily-mean data were archived for CCSM3. Transient winds $u'$ and $v'$ were generated by removing the first 12 harmonics of the annual cycle from the daily data, after which the transients were filtered using bandpass and low-pass filters given in von Storch and Zwiers (2001, columns 5 and 4 respectively of table 17.1) which retain synoptic and low-frequency timescales (roughly 2 – 6 days and longer than about 10 days). Zonal means of $u'$ and $v'$ were retained in these panels, but no significant differences were found when the fluxes were recomputed using only the zonally asymmetric components of $u'$ and $v'$.

Consistent with the sub-monthly fluxes in panel a, the synoptic momentum flux in panel d is stronger for the model than for reanalysis, only in this case the difference is much larger – up to 3 times larger for the midlatitudes. To some extent the larger momentum flux is due to stronger eddies, since the bandpass-filtered TKE in panel e is stronger for the model than for the reanalysis. However, the TKE difference is about 50% (45m$^2$s$^{-2}$ for the T42 run and 30m$^2$s$^{-2}$ for reanalysis), so the difference in strength of the synoptic eddies is small compared to the flux difference. The stronger 200mb synoptic momentum flux in CCSM3 must therefore result from a greater correlation between $u'$ and $v'$, or a stronger southwest to northeast tilt of the eddies.

North of 60$^\circ$N, the southward synoptic momentum flux is considerably larger for the model while the eddy kinetic energies are more comparable, implying a stronger northwest – southeast eddy tilt for CCSM3. In these latitudes, the synoptic contribution to zonal momentum flux is smaller than the contribution of the low-pass eddies (panel d), consistent with the idea that waves with smaller phase speeds can propagate further into latitudes of weak westerlies (e.g. Randel and Held 1991). In midlatitudes, the low-pass eddy momentum flux is in favorable agreement between CCSM3 and reanalysis, unlike the synoptic flux. But
in high latitudes (between 60° and 80° N) the southward flux of zonal momentum by low-pass eddies is stronger in CCSM3 by a factor of 2. Thus both synoptic and low-pass momentum fluxes contribute to the maintenance of the spurious Arctic surface anticyclone in the T42 CCSM3 integration.

A more thorough investigation would be required to clarify the relationship between the excessive momentum flux and the summer polar anticyclone. It would of course be difficult for the model to maintain a westerly Arctic surface circulation in the face of transients which export westerly momentum from the polar cap. Also, the midlatitude differences in fig. 13a suggest that the momentum flux bias is a bias of the general circulation rather than an Arctic phenomenon. Excess transient momentum fluxes are also found in the winter season and in the Southern Hemisphere, accompanied by strong surface zonal winds (not shown). Presumably the momentum flux biases in all seasons and regions have similar dynamical origins. On the other hand, the T85 run has the same excess northward momentum flux in midlatitudes, but weaker easterly momentum flux in the polar cap. This suggests that the midlatitude and polar cap momentum fluxes may be somewhat independent of each other. The reasons for an exclusively polar improvement at higher resolution are not clear at present.

7. Conclusions

Our examination of Arctic sea ice and atmospheric circulation consists of two parts. First, we seek to relate the spatial pattern of sea ice thickness to the seasonally varying surface winds, and account for CCSM3 thickness biases in terms of resolution-dependent surface wind biases. To isolate the effect of wind forcing we use an offline sea ice model, in which winds are specified from either reanalysis or model simulations, but all other inputs are prescribed from observational data. The offline model is used to generate an observationally derived thickness pattern, which can then be compared with thickness patterns generated in the course of CCSM3 integrations, as well as patterns generated by forcing the offline model with CCSM3 surface winds. Based on these comparisons, we find that

- The CCSM3 thickness distribution differs qualitatively from the observationally derived thickness distribution produced by the offline sea ice model. The sea ice pattern generated by observationally
derived forcings has the thickest ice along the coastline of northwestern Greenland and the Canadian Archipelago, with thickness decreasing outward across the basin in all directions (fig. 4). At both resolutions, the model produces an annual-mean pattern in which thickness is fairly uniform in a corridor stretching across the Arctic basin from the Canadian Archipelago to eastern Siberia, with a concentrated ice pile up near Wrangell Island on the Siberian side (fig. 1).

- When forced with CCSM3 surface winds, the offline model produces annual-mean sea ice patterns with thicker ice on the Siberian side than on the Canadian side, in opposition to the pattern produced by observed winds. Thus, surface wind biases alone would be sufficient to produce a reversal of the thickness gradient across the basin in CCSM3 even if the model were able to accurately simulate the observed surface temperature distribution (fig. 5).

- The reversal of the cross-polar thickness gradient produced in the offline model is consistent with surface wind differences between the NCEP/NCAR reanalysis and the CCSM3 simulations. CCSM3 surface winds are generally directed more toward Siberia and away from Canada than their reanalysis counterparts (figs. 2 and 3).

- The ice pile up around Wrangell Island in CCSM3 is more pronounced at T42 resolution than at T85. However, the T85 winds produce thicker ice along the Siberian coast in the offline model. Thus, we cannot conclude that the reduction in Siberian sea ice thickness at higher resolution is due to the mechanical effect of resolution-dependent surface wind biases. Our results suggest that differences in thermodynamic forcing must also play a role.

- Ice builds up along the Canadian coastline in the T85 integration as in the observationally derived thickness pattern (figs. 1 and 4), but no comparable Canadian maximum occurs at T42. The offline simulations show a similar Canadian thickness discrepancy in the sea ice response to surface winds from the T85 and T42 integrations. In the offline model the discrepancy can be related to the direction and strength of the spring and summer winds, which are more anticyclonic in the T42 integration, and more strongly offshore along the Canadian coastline. The largest improvement in the T85 Arctic surface winds is the disappearance of the spurious summertime Polar anticyclone of the T42 integra-
tion. This improvement in summer surface winds apparently plays an important role in establishing the T85 ice thickness maximum along the Canadian coastline.

The second part of our study attempts to relate the biases in CCSM3 Arctic surface winds to the model’s simulation of the large-scale atmospheric circulation. We examine key circulation fields which are closely related to the high-latitude Northern Hemisphere surface winds. Our examination of the structure and dynamics of the large-scale circulation reveals that

- With the exception of the summer season, Arctic SLP is too low in CCSM3 at both resolutions, and the low bias is more pronounced at the higher resolution. In particular, the Beaufort high is extremely weak in the fall, winter, and spring seasons (fig. 8).

- The three-dimensional zonally asymmetric circulation associated with the DJF Beaufort high, as revealed by plots of stationary eddy geopotential height from the NCEP/NCAR reanalysis, has an equivalent barotropic structure, with slight upward amplification and only a small intensification near the surface. Above the surface the high is distinct from the Siberian high, and the center of the high is not associated with the coldest surface or lower-tropospheric temperatures (fig. 9). In the CCSM3 integrations, the highest DJF SLP is found along the Arctic coast of eastern Siberia, with a cold-core baroclinic vertical structure associated with the coldest lower-tropospheric temperatures at that latitude. This cold-core baroclinic structure suggests that the high is an extension of the Siberian high rather than a distinct Beaufort high (fig. 9).

- Summertime (JJA) Arctic SLP is too high in the T42 integration, which has a local SLP maximum near the pole in association with the anticyclonic summer surface wind bias. The erroneous polar high is virtually eliminated in the T85 integration, as can be seen in plots of zonally averaged SLP (fig. 11).

- The observed Arctic summer circulation is characterized by a well-known North Polar low, accompanied by surface westerlies which, in the zonal mean, are comparable in strength to the midlatitude westerlies (fig. 11). In the NCEP/NCAR reanalysis, the Arctic westerlies are maintained by an Arctic Ferrel cell (fig. 12a), with northward eddy momentum fluxes in the upper troposphere (primarily by
low-frequency eddies) and a thermally indirect mean meridional circulation with rising motion at the pole. A poleward lobe of the upper-tropospheric jet extends into the domain of the Arctic Ferrel cell.

- Despite the elimination of the T42 polar SLP maximum, the T85 integration does not produce a well-defined polar low. The summertime Arctic Ferrel cell and poleward lobe of the Northern Hemisphere jet also do not appear in the T85 integration, and the surface winds do not show the coherent polar cyclone found in the reanalysis.

Our goal in presenting these findings is to offer constructive criticism, and in pursuit of that goal we have omitted a detailed discussion of the many ways in which CCSM3 is an improvement over its predecessors. For example, the thickness distribution in fig. 1 is clearly preferable to the one reported by Weatherly et al. (1998; their fig. 7a, discussed in section 3a above) for the earlier CSM, regardless of any flaws found here.

Our results suggest two goals for future model development: improved simulation of the Beaufort high, and improved simulation of the summertime North Polar low. Questions naturally arise as to whether these are in fact reasonable, attainable goals. In particular, can we be sure that a better simulation of the large-scale circulation will necessarily improve the thickness distribution? Should we expect a state-of-the-art GCM to produce a realistic circulation over a relatively small region like the Arctic? For that matter, how important is a realistic simulation of the thickness distribution?

Regarding the last question, it could be argued that the mean sea ice thickness is more important than its spatial distribution, since the mean thickness is more closely associated with the polar amplification of climate change. For example, Holland and Bitz (2003) found a close association between mean thickness and polar amplification in climate models in spite of the great variety of thickness patterns in the models they examined. Nevertheless, a realistic thickness distribution could be quite important for the export of sea ice into the North Atlantic. Most of the export occurs in the transpolar drift stream (e.g. Serreze et al. fig. 1), which transports ice preferentially from the Siberian Arctic into the North Atlantic along the west coast of Greenland. Putting the ice in the right place could thus be important for correctly simulating the export, which regulates the salinity (and hence the stability) in regions of oceanic convection and helps determine
the location of the North Atlantic sea ice edge (e.g. Bitz et al. 2005). Serreze et al. (1989) have also argued that ice production in the Canada basin, which occurs in autumn as a consequence of cyclonic summertime winds, could be important for brine production and surface heat flux.

As for the ability of coupled models to simulate the regional circulation of the Arctic, a realistic Beaufort high is certainly within reach of current models. In fact, the consensus SLP pattern in the AMIP simulations of Bitz et al. (2002) had a Beaufort high that was somewhat too strong. The problem which arises with simulations of the Beaufort high is the sensitivity of the sea ice thickness to errors in the location and strength of the high. The ice thickness bias found here is similar to the one found by Bitz et al. (2002), even though the Beaufort high in their study was too strong rather than too weak (in their case, poor placement of the high was responsible for the thickness error). Thus, a relatively high degree of accuracy in the simulation of the Beaufort high will be necessary to prevent errors in the thickness distribution. The requirements for producing such an accurate simulation are not well understood at present.

The summer circulation also poses a challenge for climate models. Since Reed and Kunkel (1960), the prevailing view has been that the polar low is a “graveyard of storms”, in which low SLP is maintained by the incursion and stagnation of cyclones from the Arctic front. Here we show that the surface circulation around the low is maintained against surface friction by transient momentum flux, a result which also emphasizes the role of the transients. It is reasonable to assume that the transients will be simulated more accurately at higher resolution, although it is not clear how much resolution is required. For example, Lynch et al.’s (2001) study of the Alaskan Arctic frontal zone found that the frequency of frontal systems depended on the representation of Alaskan topography, simulated at 30km resolution in their model. But it should also be noted that at both resolutions considered here, CCSM3 transient momentum fluxes are substantially stronger than their reanalysis counterparts throughout the northern extratropics. We believe that this bias in the transient fluxes will have to be addressed before a satisfactory simulation of the summer Arctic circulation can be achieved.
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References


——, and C. M. Bitz, 2003: Polar amplification of climate change in the coupled model intercomparison project. Climate Dynamics, 21, 221-232


FIGURE 1: Climatological sea ice thickness from NCAR CCSM3 control run simulations at T85 (left column) and T42 (right column) resolutions (see text for details). The three rows are the annual mean (top), April (middle), and September (bottom) thickness distributions. Contour interval is 0.5m, with contouring and shading suppressed when sea ice occupies less than 3% of the area of a grid box.
FIGURE 2: Climatological winds at the lowest model level ($\sigma \approx 0.992$) for CCSM3 at T85 (left column), and T42 (middle column), and the difference of T85 and T42 (right column) for winter (DJF, top row), spring (MAM, second row), summer (JJA, third row), and fall (SON, bottom row). Map domain begins at $65^\circ$N, and wind vectors are suppressed at the pole. The scale for all vectors is shown in the lower right corner of the upper left plot.
FIGURE 3: Left side: climatological surface winds from the NCEP/NCAR reanalysis for winter (DJF; a), spring (MAM; b), summer (JJA; c), and fall (SON; d). Right side: difference in climatological surface winds between CCSM3 at T85 resolution and NCEP/NCAR reanalysis (CCSM3 minus reanalysis) for winter (e), spring (f), summer (g), and fall (h). plotting conventions as in figure 2.
FIGURE 4: April (left column), September (middle column) and annual mean (right column) climatological sea ice thickness simulated by an offline sea ice model. Ice thickness patterns were generated by forcing the offline model with climatological seasonally varying surface winds from the NCEP/NCAR reanalysis. Contour interval is 0.25m.
FIGURE 5: April (left column), September (middle column) and annual mean (right column) climatological sea ice thickness simulated by an offline sea ice model. Ice thickness patterns were generated by forcing the offline model with climatological seasonally varying surface winds from the T42 (top row) and T85 (bottom row) CCSM3. Contour interval is 0.25m.
FIGURE 6: Offline ice thickness patterns generated as in figure 5, but using winds from the T42 and T85 integrations in the months March through August. Surface winds from the NCEP/NCAR reanalysis were used in the other months. Contour interval is 0.25m.
FIGURE 7: Offline ice thickness patterns generated as in figure 5, but using winds from the T42 and T85 integrations in the summer months, June through August. Surface winds from the NCEP/NCAR reanalysis were used in the other months. Contour interval is 0.25m.
FIGURE 8: Climatological SLP from the NCEP/NCAR reanalysis (top row) and CCSM3 control runs at T85 (middle row) and T42 (bottom row) resolutions for winter (DJF, left column), spring (MAM, second column from the left), summer (JJA, third column from left), and fall (SON, right column). Contour interval is 3mb, with shading for SLP in excess of 1017mb, between 1011 and 1014mb, and less than 1008mb. Map domain begins at 30°N.
FIGURE 9: DJF Eddy geopotential height at 1000mb (a) and 700mb (b) from the NCEP/NCAR reanalysis. Contour interval in both panels is 25 m, with dark (light) shading for positive (negative) values in excess of 25 m. (c) 1000—100 mb zonal-vertical cross sections of eddy geopotential height and eddy temperature at 75°N for NCEP/NCAR reanalysis. Contour interval for eddy height in c is 10m, and eddy temperature is plotted in 2K contours, with dark (light) shading for positive (negative) values in excess of 2K, and zero contours suppressed. The whitespace circle in (a) shows the 75°N latitude circle.
FIGURE 10: DJF Eddy geopotential height at 1000mb for CCSM3 at T85 (a) and T42 (b) resolution. (c) and (d): 1000–100mb zonal-vertical cross sections of eddy geopotential height and eddy temperature at 75°N for T85 (c) and T42 (d) integrations. Contours and shading as in the previous figure.
FIGURE 11: JJA zonally averaged SLP for CCSM3 at T85 and T42 resolution (a), and for the NCEP/NCAR and ERA-40 reanalyses (b), and zonally averaged surface zonal wind for the T42 and T85 CCSM3 integrations and the NCEP/NCAR reanalyses (c). In all panels, dotted (dashed) curves represent the T85 and T42 integrations, and solid lines are used for the NCEP/NCAR reanalysis. In (b) the dashed-dotted line shows the ERA-40 reanalysis.
FIGURE 12: JJA zonal-mean meridional-vertical circulation and zonal-mean zonal wind for the NCEP/NCAR reanalysis (a), the ERA-40 reanalysis (b), and CCSM3 at T85 (c) and T42 (d) resolutions. Right panels: pressure vertical velocity for the NCEP/NCAR (e), ERA-40 (f), and T85 (g) and T42 (h) CCSM3 integrations. Streamlines in (a) - (d) depict the mean meridional circulation (see text), and the zonal-mean zonal wind is depicted using 5 and 15 ms$^{-1}$ contours. In (e) - (h), the contour interval is 0.003 Pa s$^{-1}$, with dark (light) shading for positive (negative, i.e. rising motion) values in excess of 0.003 Pa s$^{-1}$. 
FIGURE 13: JJA column-mean meridional flux of zonal momentum from the NCEP/NCAR reanalysis and CCSM3 by sub-monthly transients (a) and the time-mean JJA flow (b), as described in section 6. (c) JJA column-mean sub-monthly transient kinetic energy from NCEP/NCAR reanalysis and CCSM3. (d) 200mb momentum flux by bandpass filtered transients with timescales of $\sim 2 - 6$ days (see section 6 for details). (e) Kinetic energy of transients used in panel (d). (f) 200mb momentum flux by low pass filtered transients with timescales greater than 10 days. In all panels, solid lines show the reanalysis, dashed lines show the T42 integration, and dotted lines depict the T85 integration. Units are $m^2s^{-2}$. 