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Planetary and synoptic scale adjustment of the Arctic atmosphere to sea ice cover changes

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Abstract

By means of unforced simulations with a global coupled circulation model it is shown that naturally occurring changes between high and low sea ice cover phases of the Arctic Ocean exert a strong influence on the Northern Hemisphere storm tracks. This work emphasizes the nonlinear dynamical feedback between Arctic sea ice cover and the Arctic Oscillation (AO) such as atmospheric response depending upon the wintertime sea ice distribution. Two seven year long time slices, with high and low sea ice cover, were analyzed with respect to the feedbacks between the time-mean flow, the quasi-stationary planetary and the baroclinic waves. The wave energy fluxes on time scales of 2 to 6 days increase in the middle troposphere between 30 and 60° N during the high sea ice phase and increase the zonal wind. This increase is compensated by a strong reduction in the Eliassen-Palm fluxes on time scales from 10 to 90 days between 60 and 70° N during high sea ice phases, accompanied by reduced zonal winds. High sea ice cover phases are related to the zonal wind changes during the positive phases of the AO, especially over the northern part of the Atlantic Ocean.

1. Introduction

The main causes of the interannual variability of the Arctic sea ice cover are the year-to-year variations in the atmospheric fields of the wind and temperature due to the high sensitivity of the Arctic sea ice cover to atmospheric forcing as discussed by *Arfeuille et al.* [2000]. Sea ice introduces additional feedbacks into the coupled climate system, enhancing natural variability in the polar regions with pronounced decadal trends as described by *Polyakov et al.* [2003]. *Cavalieri and Häkkinen* [2001] suggested that changes in the phase of atmospheric planetary waves are the main driver of the low-frequency variability of the ocean and sea ice in the Arctic. The decline of sea ice during the last decades attributed to global warming is, to a large extent, linked to the changes in the atmospheric circulation characterized by the North Atlantic Oscillation (NAO) and the Arctic

Oscillation (AO), which are shifted to a stronger positive phase [*Rigor et al.* 2002; *Dorn et al.* 2003; *Alexander et al.* 2004]. Years with a high NAO index are characterized by a cooling and an increased ice cover in the Beaufort Sea and a warming and a decreased ice cover in the Greenland and the Barents Seas, as shown by *Deser et al.* [2000].

Changes in sea ice concentration during winter primarily result from surface heat flux and wind stress forcing of the ice by the atmosphere [*e. g.* *Proshutinsky and Johnson*, 1997]. In the Atlantic region, strengthening of the NAO is associated with an intensification of the Icelandic Low and advection of anomalously warm air along the east coast of Greenland. As a result, ice extent increases in the Labrador Sea and decreases in the Greenland, Iceland, and Norwegian Seas, a pattern that exhibits both pronounced decadal variability and a long-term trend.

One major exception to the paradigm of the atmosphere directly forcing ice variability occurs in the Greenland Sea where the East Greenland Current transports ice southward through the Fram Strait [*Walsh and Chapman*, 1990], which can lead to large and long-lived sea ice anomalies in the North Atlantic. The coherent variability in the atmosphere-ocean-ice system in the Arctic/North Atlantic leads to a feedback loop for decadal oscillations proposed by *Mysak and Venegas* [1998], which requires the assumption that sea ice anomalies have a pronounced impact on the atmosphere. Positive ice anomalies are created in the Beaufort Sea during the positive NAO phase. These anomalies are transported out of the Arctic and produce an unusual high ice concentration in the Greenland Sea some years later. This results in a reduced winter heat flux to the atmosphere in this region leading to a reduction in the intensity of the low pressure systems in the Atlantic and modifying the NAO pattern. This negative feedback can trigger a feedback loop on a time scale of about 10 years. The formation of ice anomalies associated with the NAO and their propagation in the Arctic have been confirmed by *Arfeuille et al.* [2000].

Observational analyses suggest that sea ice anomalies affect the overlying atmosphere. *Deser et al.* [2000] found that reductions in the Greenland Sea ice cover and the associated anomalies in the

air-sea heat fluxes result in a northward shift of the local storm track. Reduced ice in the Greenland Sea during winter is associated with decreased sea level pressure and 500 hPa geopotential heights and increased surface air temperature. Although it is clear that sea ice changes influence the atmosphere, it is difficult to establish cause and effective relationship only from data analysis without model experiments as suggested by *Honda et al.* [1996].

The mean impact of the sea ice anomalies on the atmospheric circulation has been investigated by *Deser et al.* [2004], *Alexander et al.* [2004] and *Rinke et al.* [2006] but an unresolved issue is the explanation of the relative contribution of sea ice cover changes for the atmospheric response and adjustment, not only with respect to the quasi-stationary planetary waves but also concerning the synoptic storm tracks and the nonlinear feedbacks between them.

By means of simulations with an atmosphere-ocean general circulation model (AOGCM) we investigate the influence of realistic sea ice cover changes by diagnosing the nonlinear feedbacks between the Arctic sea ice cover, quasi-stationary planetary waves and synoptic storm tracks during winter. A main focus of the analysis is on the feedbacks between atmospheric conditions and sea ice cover changes and the separation between quasi-stationary planetary wave and synoptic-scale changes during the nonlinear adjustment process.

2. The global model set up

To analyze the nonlinear feedbacks between atmospheric and sea ice cover changes in the global climate system, the AOGCM ECHO-G at T30 resolution [*Zorita et al.*, 2004] has been applied with improved Arctic sea ice and snow albedo feedbacks as in *Benkel and Køltzow* [2006]. As shown by *Stendel et al.*, [2006] in experiments at T42 resolution, the representation of atmospheric blocking events depend on the model resolution and this could influence the sea ice export from the Arctic.

The model was set up with present day forcing conditions with a fixed solar constant and greenhouse gas concentrations, and integrated in time to analyze natural climate changes not influenced by time dependent changes in external forcing conditions.

Hurrell and van Loon [1997] have shown that the power spectrum of the observed NAO index has a recurring maximum at 6 to 10-year periods. Thus for the sensitivity studies of climate variability two consecutive seven year periods from a 500 year long unforced simulation have been selected. The time slices were selected on the basis of the winter (December-February) values of time-series of the total sea ice covered area (Figure 1). The first reference period is selected as a time slice representing a period of significantly larger sea ice cover than the average (hereafter the “high ice phase”). The second period is representative for smaller sea ice covered areas, with less sea ice than the average (hereafter the “low ice phase”).

3. Sea ice cover and atmospheric adjustment

The 500 hPa winter geopotential differences in Figure 2 exhibit pronounced changes over high and middle latitudes in agreement with *Deser et al.* [2004]. The lower values of geopotential height over the Arctic Ocean are connected with the higher sea ice cover. Positive geopotential anomalies are simulated over the Atlantic and Pacific Oceans during the high ice phase, in agreement with *Dethloff et al.* [2006]. The wave-like disturbances induced by the sea ice cover changes are visible in the 500 hPa zonal wind component differences between the high and low ice phase in Figure 3a. A zonally symmetric pattern with strong wave-like positive and negative wind anomalies over the Atlantic and the Pacific Oceans is excited. Figure 3b displays the zonally averaged cross section of the zonal wind component differences between high and low ice phases. An enhancement of the zonal wind by 3 ms^{-1} at mid- and high latitudes of the Northern Hemisphere is simulated. Thus high sea ice periods correspond to zonal wind changes during positive AO/NAO phases.

The horizontal heat flux at 850 hPa on synoptical time scales from 2 to 6 days for the high sea

ice phase, not shown here, corresponds to the well-known storm track pattern in the Northern Hemisphere. The corresponding heat flux differences between the high and the low ice phases at 850 hPa show a reduction of the baroclinic storm tracks over the Arctic during the high ice phase and a southward displacement over the Atlantic and Pacific as a result of the meridional diabatic heating changes. Decreased storm track activity during the high ice phase occurs especially over the southern part of Greenland and the Fram Strait.

The simulated atmospheric response of the storm tracks to sea ice cover changes is connected to anomalies in the quasi-stationary planetary wave patterns on seasonal time scales. These have been analyzed by using the zonally averaged localized Eliassen-Palm (EP) fluxes which diagnose the wave activity in the tropo- and stratosphere. The method proposed by *Trenberth* [1986] is a diagnostics of the impact of transient eddies on the time mean flow. Monitoring the energy growth in the dominant modes provides insights into the large-scale dynamical processes that control the model response to changes in the Arctic sea ice cover.

Figure 4a shows the additional eddy heat and momentum fluxes of the troposphere expressed by the zonally averaged EP flux differences between the high and low sea ice phase on all time scales from 2 to 90 days. A pronounced maximum in the latitudinal belt 30-60° N appears in the lower and middle troposphere and a second maximum in the upper troposphere at 200 hPa and 30° N. A remote influence in the mid-troposphere of the Southern Hemisphere can be seen.

Figure 4b displays the zonally averaged EP flux differences between the high and low sea ice phase on synoptical time scales 2 to 6 days. The maxima described in Fig. 4a are mainly a result of baroclinic wave activity. Figure 4c presents the zonally averaged EP flux differences between the high and low sea ice phases on seasonal time scales 10 to 90 days. The EP fluxes due to quasi-stationary planetary waves are reduced in the high sea ice phase and opposite to those induced by baroclinic waves. The strongest reduction occurs between 900 and 500 hPa in the latitudinal belt 60-80° N.

Compared to the zonally averaged wind differences in Figure 3b, the zonal wind increase between 50 and 60° N is due to the influence of EP fluxes on synoptical time scales, whereas the wind reduction between 60 and 70° N is a result of the EP fluxes due to quasi-stationary waves on seasonal time scales. These results show that the adjustment of the atmospheric response to high and low sea ice phases takes place between the time averaged zonal wind, the quasi-stationary waves and the baroclinic storm tracks. During the high sea ice phase the zonal mean wind between 30 and 60° N and the synoptic EP fluxes are increased but the quasi-stationary EP fluxes north of 60° N are reduced. Sea ice and snow cover changes force significant anomalies in atmospheric dynamical fields connecting the Arctic and tropical-mid latitudes through planetary wave guides and lead to readjusted feedbacks between the time mean zonal flow, planetary waves on seasonal time scales and baroclinic waves on time scales from 2 to 6 days with opposed responses in the EP fluxes.

The contribution of quasi-stationary planetary waves to the vertical EP fluxes is stronger in the low ice phase as a result of the changed thermal forcing gradients between the ice covered ocean and the ice free ocean. The baroclinic waves behave opposite and respond with increased EP fluxes during the high sea ice phase.

The atmospheric response to the sea ice state and the triggering of the positive AO/NAO are determined by sensitive nonlinear feedbacks between the synoptical waves and planetary waves on seasonal time scales.

4. Summary and conclusions

Naturally occurring changes of the Arctic sea ice cover between high and low ice phases exert a strong influence on the mid- and high-latitude climate by modulating the strength of the sub-polar westerlies, the quasi-stationary planetary waves and the baroclinic storm tracks. The atmospheric planetary and synoptic scale response to sea ice cover changes moves in the opposite direction and

builds a very sensitive nonlinear feedback system under the influence of different amounts of sea ice cover. A pronounced maximum in the EP fluxes due to baroclinic waves occur between 900 and 500 hPa in the latitudinal belt 30-70° N. A second weaker maximum exists also in the upper troposphere at 200 hPa and 30° N.

During high sea ice phase, the zonal wind north of 40° N increases and is reduced southward. The wave energy fluxes on time scales 2 to 6 days increase in the middle troposphere between 30 and 60° N during the high sea ice phase forcing the increased zonal wind. Their increase is compensated by a strong reduction in the Eliassen-Palm fluxes on time scales from 10 to 90 days between 60 and 70° N, accompanied by reduced zonal west winds. High sea ice cover phases resemble zonal wind changes during the positive phases of the AO. Regional sea ice or snow anomalies in the Arctic can influence the quasi-stationary planetary waves and transient storm tracks on synoptical to seasonal scales. The zonal wind difference pattern between high and low sea ice phases reveals similarities with that of the positive AO pattern and reveals a hint that high sea ice phases could trigger an enhanced positive AO-like atmospheric circulation pattern.

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Figure caption

1 **Figure 1.** High and low phases of Arctic sea ice cover ($\times 10^6 \text{ km}^2$) for winter (DJF) (top) and annual
2 mean (bottom) in the ECHO-G model.

3 **Figure 2.** Geopotential height differences (gpm) between high (7 yrs) and low (7yrs) ice phases at
4 500 hPa, for winter (DJF).

5 **Figure 3.** Zonal wind component differences (ms^{-1}) between high (7 yrs) and low (7 yrs) ice
6 phases, for winter (DJF) (a) at 500 hPa, (b) zonally averaged latitude-height cross section.

7 **Figure 4.** Zonally averaged EP flux differences ($10^6 \text{ m}^3\text{s}^{-2}$) between high (7yrs) and low (7 yrs) ice
8 phases for winter (DJF) for (a) all time scales (2-90 days), (b) baroclinic time scales (2-6 days), (c)
9 seasonal time scales (10-90 days). Colours display the magnitude of the differences, the arrows
10 describe the differences in the EP vector propagation. describe the differences in the EP vector
11 propagation.

Figure 1

Figure 2

Figure 3 a-b

Figure 4 a –c