



## A review on Northern Hemisphere sea-ice, storminess and the North Atlantic Oscillation: Observations and projected changes

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### ABSTRACT

The Arctic has undergone substantial changes over the last few decades in various cryospheric and derivative systems and processes. Of these, the Arctic sea ice regime has seen some of the most rapid change and is one of the most visible markers of Arctic change outside the scientific community. This has drawn considerable attention not only from the natural sciences, but increasingly, from the political and commercial sectors as they begin to grapple with the problems and opportunities that are being presented. The possible impacts of past and projected changes in Arctic sea ice, especially as it relates to climatic response, are of particular interest and have been the subject of increasing research activity. A review of the current knowledge of the role of sea ice in the climate system is therefore timely. We present a review that examines both the current state of understanding, as regards the impacts of sea-ice loss observed to date, and climate model projections, to highlight hypothesised future changes and impacts on storm tracks and the North Atlantic Oscillation. Within the broad climate-system perspective, the topics of storminess and large-scale variability will be specifically considered. We then consider larger-scale impacts on the climatic system by reviewing studies that have focused on the interaction between sea-ice extent and the North Atlantic Oscillation. Finally, an overview of the representation of these topics in the literature in the context of IPCC climate projections is presented. While most agree on the direction of Arctic sea-ice change, the rates amongst the various projections vary greatly. Similarly, the response of storm tracks and climate variability are uncertain, exacerbated possibly by the influence of other factors. A variety of scientific papers on the relationship between sea-ice changes and atmospheric variability have brought to light important aspects of this complex topic. Examples are an overall reduction in the number of Arctic winter storms, a northward shift of mid-latitude winter storms in the Pacific and a delayed negative NAO-like response in autumn/winter to a reduced Arctic sea-ice cover (at least in some months). This review paper discusses this research and the disagreements, bringing about a fresh perspective on this issue.

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## 1. Introduction

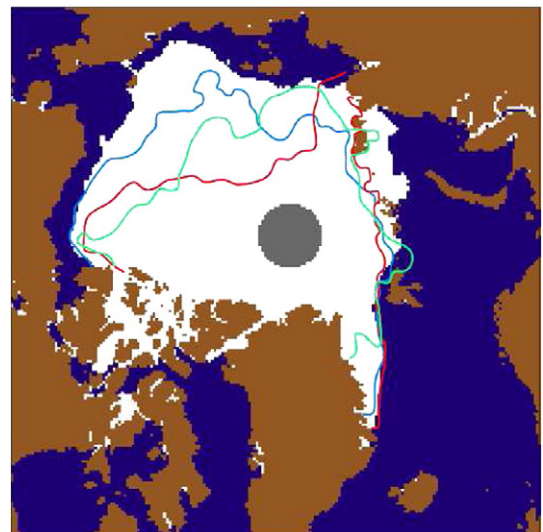
The Arctic Ocean is transforming to a seasonally ice covered state, with a reduction in the amount of generally thicker perennial ice observed since the mid 1980s (Maslanik et al., 2007; Nghiem et al., 2007). The summer minimum sea-ice extent, observed over the period of modern satellite observations of 1979–2006, has decreased by 8.6% per decade (Serreze et al., 2007, see Table 1), followed by an abrupt drop to a new record low of  $\approx 4.3 \cdot 10^6 \text{ km}^2$  in September 2007 (Stroeve et al., 2008, see our Figs. 1 (coverage), 2 (concentration trend), and 3 (extent anomalies)). Whether further decreases will be gradual or abrupt is unclear. Observations indicate that the sea-ice cover is changing significantly faster than anticipated by the modelling studies generated for the

Intergovernmental Panel for Climate Change Fourth Assessment Report (IPCC, 2007; Stroeve et al., 2007, see our Fig. 4), even if the unusual (five standard deviations anomaly) 2007 and subsequent years are disregarded. In contrast, the

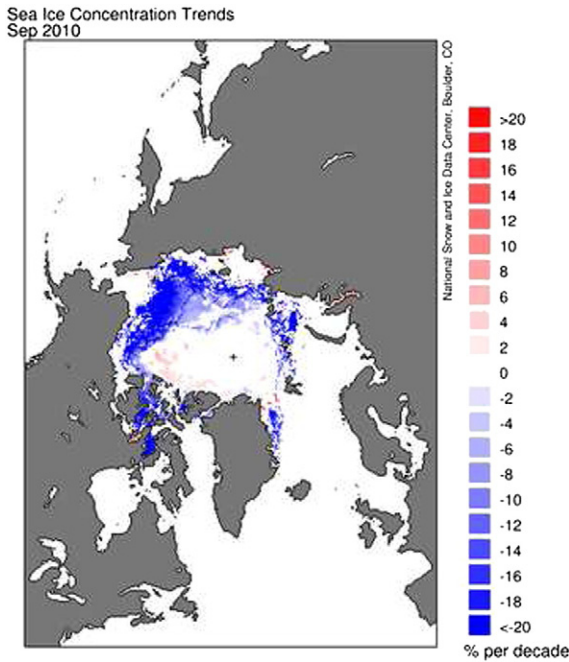
**Table 1**

Yearly, March and September average Arctic ice extent, standard deviation and trend for 1979–2006, with the 1979–2000 period used as reference to calculate the trends. The final column indicates the number of years of data necessary to detect a significant trend at the 95% confidence level. Data is taken from Meier et al. (2007).

Period	Average, 1979–2000	Std dev., $\sigma$ , 1979–2000	Trend	Number of years for significance
	$10^6 \text{ km}^2$	$10^6 \text{ km}^2$	$\% \text{ decade}^{-1}$	
Annual	12.2	0.29	−3.6	6.1
March	15.8	0.34	−2.5	7.1
Sept	7.1	0.54	−8.4	5.9



**Fig. 1.** September Arctic sea ice coverage. The median ice cover (1979–2000) is presented in solid white. The blue, green and red lines indicate the ice edges in 2005, 2008, and 2007, respectively. From Simmonds and Keay (2009).



**Fig. 2.** Arctic September sea ice concentration trend (% per decade) from observations based on the period 1979 to 2000. NSIDC (2010) figure.

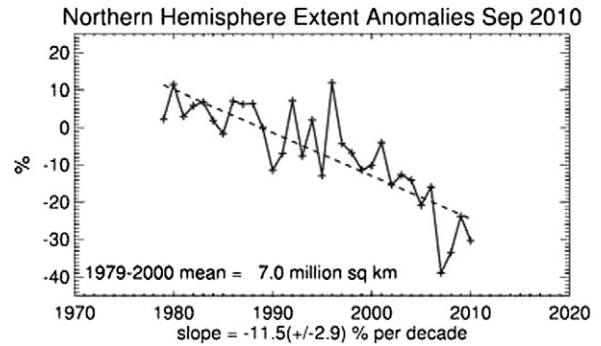
Antarctic sea-ice extent has increased and the bulk of the Antarctic has experienced little surface temperature change over the last 50 years (Turner et al., 2007; Turner and Overland, 2009).

The recent and future projected reductions in the summer Arctic ice cover are not homogeneous temporally or geographically. For example, the region north of Canada, and in particular the Canadian Arctic Islands, are essentially a refuge for multi-year ice and severe ice conditions due to dynamical deformation of first-year ice. Dynamical deformation can create features up to 20 times thicker than possible through thermodynamics – i.e., freezing – alone (Melling, 2008). This is thus a challenging area for offshore oil and gas exploration and operations. Multi-year ice is a principal disincentive for offshore oil and gas development in the Arctic and for transport through it.

The major thinning observed since the 1990s, is expected to continue in the future and to make the sea-ice cover more variable, and hence more sensitive to atmospheric forcing. Storms produce extensive fields of ridged ice. A change in storminess in a warmer climate would then affect ridging with consequences to the maximum ice thickness.

In summary, Arctic sea ice has now entered a state of being particularly vulnerable to anomalous atmospheric forcing, both in terms of dynamics associated with faster ice drift and export and thermodynamic feedbacks. A new observational study (Serreze et al., 2009) of the ice-albedo feedback strongly suggests that the long-theorised Arctic amplification<sup>1</sup> is indeed emerging, largely driven by loss of the sea ice

<sup>1</sup> Polar or Arctic amplification refers to greater temperature changes in the Arctic compared to the earth as a whole.



**Fig. 3.** Northern Hemisphere September sea ice extent anomalies (% per decade) from observations for the period 1979 to 2010. NSIDC (2010) figure.

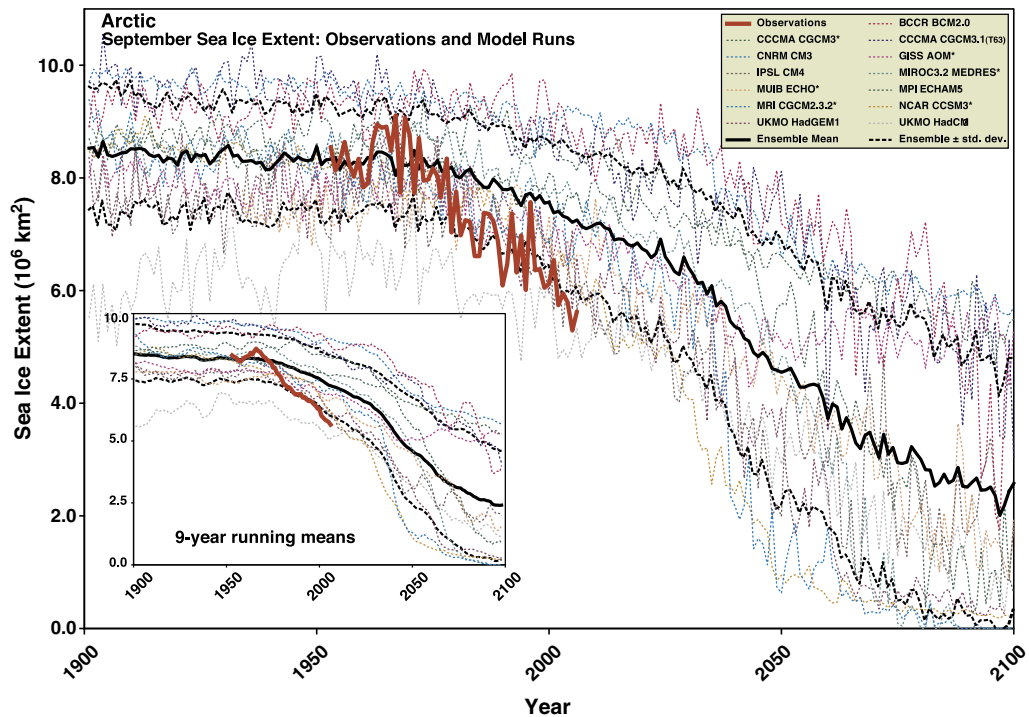
cover, allowing for strong heat transfer from the ocean to the atmosphere (see also Kumar et al., 2010; Screen and Simmonds, 2010a,b). Consistent with observed reductions in sea ice concentration and extent, fields from both the NCEP/NCAR<sup>2</sup> and regional data point to the emergence of a surface-based Arctic amplification in the last 5–10 years (Serreze et al., 2009). Starting in the late 1990s and relative to 1979–2007, the Arctic Ocean surface air temperature anomalies in the NCEP reanalysis turned positive in autumn and have subsequently grown.

Far more studies have investigated the impact of sea-surface temperatures (SSTs) (see e.g. Kushnir et al., 2002) rather than sea ice on the atmosphere. The recent Arctic sea-ice changes has, however, motivated an investigation of the impact of sea ice on the atmosphere (Alexander et al., 2004; Deser et al., 2004; Magnusdottir et al., 2004; Seierstad and Bader, 2009; Mesquita et al., 2010a). This has shown that local changes in albedo, heat fluxes and baroclinicity due to the changed sea-ice cover might affect storms travelling or developing over the sea-ice region. In addition, there are some indications that changes in sea ice might impact large-scale circulation patterns and therefore mid-latitude storms, precipitation, temperature and winds. Oceanographic observations show an increased rate of melting of sea ice in the Arctic Ocean, and the advance of an anomalously warm tongue of Atlantic water intruding across the Arctic below the halocline over the past few decades. Turner (2010) finds that vertical double-diffusive convection between the intruding warm Atlantic layer and the surface could contribute substantially to the observed increased rate of melting, but that the direct input of heat from the atmosphere leads to much larger melting rates.

In this review paper, we will summarise the current literature on the impact of the Arctic sea-ice on the atmosphere – from the perspective of storm tracks and the North Atlantic Oscillation (NAO). More specifically the paper will discuss:

- the observed and projected changes in Arctic sea ice, Northern Hemisphere (NH) extratropical storms and the NAO;

<sup>2</sup> National Centers for Environmental Prediction (NCEP)/The National Center for Atmospheric Research (NCAR).



**Fig. 4.** Arctic September sea ice extent ( $\times 10^6 \text{ km}^2$ ) from observations (thick red line) and 13 IPCC AR4 climate models, together with the multi-model ensemble mean (solid black line) and standard deviation (dotted black line). Models with more than one ensemble member are indicated with an asterisk. Inset shows 9-year running means.

From Stroeve et al. (2007).

- a comparison of recent findings from idealised studies about the impact of the loss of Arctic sea ice on storms and the NAO;
- storm tracks in the future in reduced sea-ice scenarios
- whether the recent observed and projected future decline in Arctic sea ice is/will be a major driver for the projected changes in storms and of the NAO – this will be discussed in the context of whether other forcing factors are more relevant than the impact of the reduced Arctic sea-ice cover;

## 2. Arctic sea ice

### 2.1. Arctic sea ice: a brief introduction

The most obvious characteristic feature of the Arctic Ocean is its floating sea-ice cover (Serreze et al., 2007). The thickness of sea ice is a consequence of past growth, melt and deformation. Therefore, two distinct mechanisms govern the evolution of sea ice (Perovich and Richter-Menge, 2009):

- Thermodynamics; melting and freezing of sea ice are determined by the energy fluxes among the ocean, atmosphere and sea ice. Sea ice cools and grows when the net energy flux is negative, and warms and melts when it is positive. The surface heat budget is made up of the radiative fluxes (solar radiation and longwave radiation), the turbulent fluxes of sensible and latent heat plus the heat conduction through the ice. At the ice bottom, the

heat balance consists of heat conduction through the ice and a flux of heat from the ocean to the ice. The surface and bottom ice heat budget depends on snow depth, ice thickness, albedo and ice topography.

- Dynamics; motion of sea ice, which is mainly driven by surface winds, can cause rapid changes in the thickness of the sea ice. Divergence of sea-ice cover can result in open water, while convergence causes ridging and an increase in sea-ice thickness. The acceleration of the sea-ice cover depends on surface wind stress, bottom ocean stress, sea surface tilt, the Coriolis force, and the internal ice stress. Extensive fields of ridged ice are produced by storms.

Sea-ice concentration (fraction of the ocean covered by sea ice) and extent (area enclosed by the ice edge – operationally defined as the 15% concentration contour) are the most important measures. The ice extent, or ice edge position, is the only sea-ice variable with observations available for longer than a few decades. The expansion or retreat of the ice edge may be amplified by the ice-albedo feedback (see below).

The relationship between sea ice and the atmosphere is twofold. On the one hand, the atmosphere impacts the sea-ice concentration and extent through dynamical processes: wind, storms (Simmonds and Keay, 2009), large-scale variability (Hilmer and Jung, 2000); and through thermodynamic processes: changes in turbulent energy fluxes, air temperature and radiative effects (Budyko, 1969; Sellers, 1969; Manabe and Stouffer, 1980; Gregory et al., 2002;

Stroeve et al., 2007; Serreze et al., 2007). On the other hand, sea-ice changes feed back on the atmosphere due to changes in, for example, albedo and heat fluxes. This feedback from the sea ice on the atmosphere will be the main focus of this review paper.

It has also been suggested that melting sea ice could contribute to injecting freshwater into the Arctic and Nordic seas (Curry and Mauritzen, 2005). This could impact the North Atlantic Meridional Overturning Circulation (MOC) and in turn the Atlantic storm track via local changes in the baroclinicity.

Only more recent studies have considered the direct influence of sea ice on the atmospheric variability and storms. The impact of sea surface temperatures (SST), especially SST anomalies in midlatitudes, on the atmosphere have been extensively studied in the past (Inatsu et al., 2002; Kushnir et al., 2002; Magnusdottir et al., 2004; Grossmann and Klotzbach, 2009; Hodson et al., 2010; Kwon et al., 2010). The impact of sea ice has received greater attention only in recent years (Kvamstø et al., 2004; Magnusdottir et al., 2004; Seierstad and Bader, 2009). This has been motivated by the observed strong reduction in sea ice and better observations in the satellite era. These observations have also allowed the use of data over several decades to study the relationship between sea ice changes and cyclone variability (Deser et al., 2000).

Why does Arctic sea ice matter? It matters because of four main processes: albedo feedback, heat fluxes, freshwater fluxes and “ice-edge” baroclinicity. These are summarised below.

### 2.1.1. Surface albedo

Sea ice extent and its snow cover modify the surface albedo – a measure of how strongly incoming solar radiation is reflected from the earth's surface. Ice and snow have a much higher albedo than other surface coverings, especially the ocean. Typical values of ocean albedo at high latitudes are 10%, whereas a typical sea-ice albedo with snow cover is approximately 60% (Hartmann, 1994).

The strong contrast between ocean and sea-ice albedo produces a positive climate – ice-albedo – feedback that is as follows: an increase in sea-ice cover enhances surface albedo, which decreases the amount of solar energy absorbed. This leads to a cooling, which is further enhanced by reduced ocean–atmosphere fluxes, and an increase in sea-ice cover, closing the feedback loop. The opposite works for reduction in sea-ice cover. The potential importance of ice albedo to the climate system has been pointed out from model studies (Budyko, 1969; Sellers, 1969; Shine and Henderson-Sellers, 1985; Ingram et al., 1989). In particular, for the southern parts of the marginal sea ice, such as the Sea of Okhotsk, the sea-ice albedo is expected to be a significant parameter – since solar radiation is abundant even in mid-winter (Toyota et al., 1999). Changes in sea-ice concentration can also affect the atmospheric circulation through changes in surface albedo; this affect is suppressed in winter.

### 2.1.2. Heat fluxes

Sea-ice cover reduces the sensible and latent heat fluxes to the cold atmosphere from a relatively warm ocean. Thus, it alters significantly the radiation budget (Johnson, 1980;

Walsh and Johnson, 1979). Changes in sea-ice concentration can affect the atmospheric circulation through changes in surface fluxes of heat, momentum and moisture (Deser et al., 2000). The atmospheric circulation can be sensitive to the heat fluxes associated with changes in the sea-ice cover, as for example, in the stormy region east of Greenland. The heat fluxes associated with sea-ice changes can be an order of magnitude higher than mid-latitude SST fluxes. Changes in Greenland sea-ice cover induced by the large-scale circulation may feed back upon the atmosphere by changing the cyclone activity locally.

Previous studies have shed some light on understanding the impacts sea ice can have on the atmosphere. For example, Royer et al. (1990) found a high-latitude atmospheric warming due to the changes in upward fluxes and longwave radiation in an ice-free experiment. The warming also led to a reduction in the surface pressure and a precipitation increase over the area – this could point to the changes in the storm tracks over the region as well. The results of Royer et al. (1990) are similar to previous low resolution experiments (Fletcher et al., 1973; Newson, 1973; Warshaw and Rapp, 1973).

### 2.1.3. Local baroclinicity

The presence of an ice edge produces large horizontal gradients in heat fluxes (Overland and Pease, 1982). Thus, the presence or absence of sea ice may contribute to local temperature gradients. As the sea-ice edge moves, the temperature gradient changes, but so also does the static stability so that changes in baroclinicity are the result of a complex combination of these. These changes could provide conditions that feed low-level baroclinicity, which can be important for the development of intense mesoscale polar lows. On the other hand an increase in static stability can act to suppress their formation (Zahn and von Storch, 2010).

## 2.2. Arctic sea ice observations

Observations of Arctic Ocean sea-ice<sup>3</sup> cover exist back to the 1600s, made by explorers, sealers, whalers and fishermen. Willem Barentsz obtained information on the sea ice conditions from the northern Barents Sea in 1596 when he was searching for the North-East Passage (De Veer, 1609). Information on the sea ice condition from the Eastern Barents Sea was obtained from the Dutch trading connection with the Archangel region. Observations became more frequent when whaling and sealing started in the 17th century, and since 1730 annual information on the sea ice conditions are available for most years (Vinje, 2001). Frequent observations from aircraft started around 1950 and continued until satellite based observations took over from the mid 1970s (ACSYS, 2003; Divine and Dick, 2006). Observations of sea-ice drift in the Arctic Ocean can be obtained with data from the International Arctic Buoy Programme<sup>4</sup> (IABP<sup>5</sup>).

<sup>3</sup> A nice overview is given in Thomas and Dieckmann (2003).

<sup>4</sup> See <http://IABP.apl.washington.edu>.

<sup>5</sup> IABP is a network of drifting buoys in the Arctic Ocean providing meteorological and oceanographic data for real-time operational requirements and research purposes including support to the World Climate Research Programme (WCRP) and the World Weather Watch (WWW) Programme.

The Fram Strait is the main gateway for Arctic ice export and is the most concentrated meridional ice flow in the World Ocean. This flux thus provides a measure of the net ice production in the Arctic Ocean. Since 1990, the ice thickness and the ice transport through the Fram Strait has been monitored by means of sub-sea moorings.<sup>6</sup> Also, observations of multi-year ice that originated from the Arctic Ocean travelling via the Fram Strait are available from southwest Greenland waters since 1820 (Schmith and Hansen, 2003).

The Arctic sea-ice cover has a strong seasonal cycle with a maximum in March at the end of the winter season. The sea ice shrinks during spring and summer, reaching its minimum extent in September. The observed climatological (1979–2000) Arctic sea-ice extent varies from a minimum of about  $7 \times 10^6 \text{ km}^2$  in September at the end of the melting season and a maximum of about  $15\text{--}16 \times 10^6 \text{ km}^2$  in March (Serreze et al., 2007; NSIDC, 2010, see also Table 1). In summer, this corresponds roughly to the area of the United States. In winter, between 5 and 10% of the Northern Hemisphere ocean is covered with sea ice. In March, the sea ice extends down the western side of the ocean basins reaching the Gulf of St Lawrence in the Atlantic sector and the Sea of Okhotsk in the Pacific sector. The Gulf of Bo Hai, located off the east coast of China at 38 N, is the most southerly region where a considerable sea-ice cover occurs. In contrast the Norwegian coast up to 70 N normally stays ice free.

The cause of the asymmetric sea-ice distribution between the east and west coasts of the Atlantic and Pacific basins are ocean currents and winds. In September, the Arctic sea ice is restricted to the central Arctic Ocean, with minor extensions into the Canadian Arctic Archipelago and along the east coast of Greenland (Weeks, 2002; NSIDC, 2010). The ice cover can be mainly classified as a permanent sea ice region, where the sea ice is present throughout the year, and a seasonal sea-ice area (Weeks and Ackley, 1986; Polyak et al., 2010). A substantial part of the Arctic sea-ice cover is permanent (Polyak et al., 2010).

The Arctic sea-ice extent undergoes large interannual variations. The monthly interannual average extent can vary by up to 1 million  $\text{km}^2$  (NSIDC, 2010). Compared to the Antarctic, Arctic sea ice is thicker (NSIDC, 2010), because it is less mobile due to the different geography. The Arctic Ocean is surrounded almost entirely by land except for the sector between about 20E and 20 W (Serreze, 2002). Therefore, sea ice tends to stay in the Arctic waters and is more exposed to convergence and piling up into ridges.

The typical sea-ice thickness in the Arctic is about 2–3 m (Williams et al., 1975; NSIDC, 2010). Sea ice can be roughly divided into first-year ice and multi-year ice (Nghiem et al., 2007). The first-year ice denotes only a single year growth, whereas multi-year ice has persisted for more than one melt season (Weeks and Ackley, 1986; Polyak et al., 2010). New sea ice develops in open water during autumn and winter. It is transported to the central Arctic and grows from the bottom (Polyak et al., 2010). The Arctic sea-ice cover is more or less in constant motion due to winds and ocean currents. The clockwise Beaufort Gyre in the Canada Basin and the

mean sea-ice drift from the Siberian coast, across the pole and through Fram Strait define the large-scale mean annual drift in the Arctic (Serreze, 2002).

Changes in the Arctic seasonal ice zone have been documented by Kinnard et al. (2008) using historical and satellite observations. The maximum sea-ice extent was relatively stable until the 1960s. After that there has been a gradual decrease in maximum sea-ice extent. The minimum observed Arctic sea-ice extent is more variable on interannual and decadal time-scales than the maximum extent. A more pronounced declining trend – than that of the maximum sea-ice extent – is observed after the early 1950s with an increasing rate in the last decade. They find a gradual increase in the seasonal ice zone with a marked acceleration over the past three decades. The primary reduction in the overall Arctic sea-ice thickness over the observational period from 2003 through 2008 is attributable to the thinning ( $\approx 0.6 \text{ m}$ ) of the multi-year ice cover (Kwok et al., 2009).

In a modelling study of the 20th century Arctic Ocean/sea ice system, Kauker et al. (2008) hypothesise that the century long observed declining trend in Arctic sea-ice extent is anthropogenically forced. They limit their conclusion by warning that even time series of five to six decades in length might be strongly influenced by natural climate variability. According to the US National Snow and Ice Data Center there is a significant sea-ice extent decrease of 4.1% per decade during the satellite era (1979–2008). Perhaps one of the most striking observational changes since the IPCC AR4 is the summer minimum sea ice extent recorded in 2007 that was not predicted by climate models (Allison et al., 2009). Due to internal climate variability it should not be a surprise that forced climate models do not simulate all the details of climate evolution (see also Deser et al., in press).

### 2.3. Projected arctic sea-ice changes

Wang and Overland (2009) used a set of multi-model output and made an offset-adjustment based on the 2007/2008 September minima in the Arctic sea ice to predict a nearly sea-ice-free ( $< 10^6 \text{ km}^2$ ) Arctic in September by the year 2037. This study was based on several CMIP3 simulations and the estimated number of years for sea-ice extent to drop from the current observed value to less than 1 million  $\text{km}^2$ . They find that anthropogenic forcing is a necessary condition for a major future sea-ice loss to occur. However, natural variability in the form of both recent warm years and wind-driven sea-ice drift could explain the observed minimum of September 2007 sea-ice that occurred many decades earlier than expected due to the influence of greenhouse gas forcing alone.

The IPCC AR4 models vary widely in the simulated reduction of sea ice for the recent decades and those projected for the 21st century. None or very few individual model simulations show trends comparable to observations – in particular, they underestimate the observed trend since the 1950s (Stroeve et al., 2007). About half of the IPCC AR4 models forced by the Special Report on Emissions Scenarios (SRES) A1B – in which atmospheric  $\text{CO}_2$  approximately doubles by 2100 – simulate an ice-free summer Arctic by 2100 (Arzel et al., 2006). By contrast, even by the late 21st century, most

<sup>6</sup> The moorings are instrumented with upward looking sensors to measure the draught and Doppler current profiler to measure the speed of the sea ice (Widell et al., 2003).

models project a thin Arctic sea-ice cover in March (Serreze et al., 2007).

There is clearly large uncertainty in simulating the present day and future Arctic sea-ice cover and thickness. Despite the large differences among individual models, the multi-model average sea ice extent agrees reasonably well with the observations for the period 1981–2000 (Arzel et al., 2006). The spread among models is due to several causes, such as the initial (late-20th century) simulated ice state, the radiative forcing, the parametrisation of the effect of surface melt on the absorption of short wave radiation, the modelled ocean circulation, simulated cloud conditions, and natural variability in the modelled system (Laxon et al., 2003; Serreze et al., 2007).

Holland et al. (2006) find abrupt changes in the summer Arctic sea ice in 21st century projections from the Community Climate System Model version 3 (CCSM3). As sea ice thins due to global warming, it becomes more vulnerable to natural climate variability as it melts more easily. The melting is further amplified by the ice-albedo feedback. In the events simulated by CCSM3, anomalous ocean heat transport acts as a trigger. Such abrupt transitions are typically four times as fast as the observed trends over the satellite record. In one ensemble member this abrupt change results in near ice-free September conditions by 2040. A number of other climate models show similar rapid ice loss events. The future emission scenario used to force the model affects the likelihood of abrupt sea-ice reductions.

The study of Flato and Participating CMIP Modeling Groups (2004) shows that the basic state of the sea ice and the reduction in thickness and/or extent have little to do with sea ice model physics among the CMIP2 models. Holland and Bitz (2003) (CMIP2) and Arzel et al. (2006) (CMIP3) find serious biases in the basic state of simulated sea ice thickness and extent. Further, Rind et al. (1995) ( $2\times CO_2$ ), Holland and Bitz (2003) (CMIP2) and Flato and Participating CMIP Modeling Groups (2004) (CMIP2) show that the initial state of the sea ice thickness and extent have a significant influence on the projected change in sea-ice thickness in the Arctic and extent in the Antarctic.

### 3. Storms

#### 3.1. Storms and storm tracks

The centre of a closed surface cyclonic (counter-clockwise in the Northern Hemisphere) circulation outside of the tropics is normally referred to as an extratropical cyclone. A strict definition of when an extratropical cyclone is called a storm is when the wind speed attains values greater than 24.5 m/s (after the Beaufort Wind Scale, see WMO, 1970). Storms can last anywhere from between 12 and 200 h, depending on season and geography and can vary in size from the mesoscale ( $\leq 1000$  km) to synoptic scale (1000 km). Storms are often associated with damaging winds (Mesquita et al., 2009) and/or strong precipitation in the form of rain and snow and are an integral part of the general circulation of the atmosphere transporting heat, moisture (Sorteberg and Walsh, 2008) and momentum polewards.

Originally, the term storm track referred to the tracks of individual cyclonic weather systems, but now more often

refers to regions where synoptic-scale storms prefer to travel. In the Northern Hemisphere, the main storm corridors are in the North Pacific and North Atlantic Oceans (Hoskins and Hodges, 2002; IPCC, 2007). This is mainly due to the distribution of ocean and land in the Northern Hemisphere and the associated meridional temperature gradients. High latitude mesoscale cyclones, generally referred to as polar lows, are also of importance associated with damaging weather (Zahn and von Storch, 2010; Kolstad and Bracegirdle, 2008).

An important event in meteorology was the first formalised description of the life-cycle of a mid-latitude cyclone (Bjerknes and Solberg, 1922). The baroclinic instability<sup>7</sup> framework has since been developed and extensively explored to explain the development and life-cycles of extratropical storms. A more mathematical description was first supplied by Eady (1949), Charney (1947), and Charney and Stern (1962) who formulated quasi-geostrophic models that described the 3D structure of baroclinic growth of finite amplitude perturbations to the zonal flow. Since then, alternative conceptual and mathematical models have been developed.

There are two leading conceptual models describing cyclogenesis: In the first, an upper-level potential vorticity (PV) anomaly travels over a region with a horizontal surface temperature gradient. This configuration can induce a low-level PV anomaly which in turn will reinforce the upper-level one, making it grow and become unstable; in the second, upper and lower Rossby wave interact (lock), become unstable and grow. This is the counter propagating Rossby-wave perspective (Heifetz et al., 2004). The origins of these two models actually go back further (Davies, 2010). More recent descriptive conceptual models are those of Shapiro and Keyser (1990) for oceanic cyclones, where the configuration of the fronts is somewhat different from the Norwegian Cyclone Model (see e.g. Martin, 2006), and that of Browning (1986), who described the flows inside cyclones in terms of the warm and cold conveyor belts and the region of dry descending air behind the cold front called the dry slot. Whilst the baroclinic theory can describe synoptic scale cyclones well, meso-cyclones can have a more diverse range of development mechanisms including the influence of convective processes, e.g. polar lows (Rasmussen and Turner, 2003).

Storms are not only tied to temperature gradients for their development, but may also be formed due to orographic effects where vorticity can be created or destroyed as vortices are stretched or compressed by the diverging or converging isentropes (Holton, 2004). For example, the Rocky Mountains serve as an important geographic feature for the generation of Northern Hemisphere storm tracks. In addition, diabatic heating can also be important for their development and intensification (Chang et al., 1982; Hoskins and Valdes, 1990; Posselt et al., 2008).

The North Pacific and North Atlantic storm tracks may be considered as self-maintaining, since the diabatic heating maxima in the storm track region are caused by horizontal

<sup>7</sup> A measure of baroclinic instability is the Eady growth rate. Simmonds and Lim (2009) have shown that caution must be used when calculating Eady growth rate directly from the time-mean flow to diagnose changes in cyclone properties.

and vertical flow displacements, which are in turn a by-product of the individual storms (Hoskins and Valdes, 1990). Storm tracks can also be self-maintained by the fact that they help drive the Kuroshio current due to the low-level mean flow induced by the eddy activity, and in turn, the Kuroshio current (strong baroclinicity) is crucial for the existence of the North Pacific storms (Hoskins and Valdes, 1990). Other results have shown that storm tracks are not completely self-maintained, but that planetary stationary waves are crucial for organising them (Broccoli and Manabe, 1992; Chang et al., 2002; Inatsu and Hoskins, 2004).

Storm track transients may also be considered as an ensemble of wave packets (with wave growth and decay) happening over all portions of the storm track (Chang and Orlanski, 1993). The individual synoptic eddies within the nonlinear wave packets decay by transferring their energy to their neighbour eddies downstream (downstream development). Storm tracks are also closely linked to the large-scale flow (Pezza et al., 2010). A storm track shift caused by SST and sea-ice changes has often a larger impact on the large-scale flow than the direct impact of the imposed SST/sea-ice forcing (Held et al., 1989; Cai and Mak, 1990; Chang et al., 2002). Storm tracks are also interconnected, in the sense that each track may seed its downstream neighbours in the form of upper-level troughs around the globe (Chang et al., 2002; Pierrehumbert, 1986). There is also a robust positive correlation in the variation between the Pacific and Atlantic storm tracks, where eddies in the Atlantic storm track are seeded by the Pacific track (Chang and Fu, 2002). A stronger Pacific storm track can lead to a stronger Atlantic storm track (Chang and Fu, 2002; Strong and Magnusdottir, 2008).

### 3.1.1. Storm track diagnostic techniques

There are two main approaches to producing diagnostics of the storm tracks: Eulerian and Lagrangian. The most common Eulerian method is the bandpass filtered variance, typically in the synoptic 2–6 day band, and in this context referred to as baroclinic waveguides (Blackmon, 1976; Blackmon et al., 1977). These waveguides extend across the Atlantic and the Pacific, and are often called “storm tracks” in analogy to the Lagrangian perspective (Chang et al., 2002). The limitations of this approach are that even though the filtered variance coincides with the major storm track regions, it can only provide a fairly general indication of storm track activity, as it does not discriminate between cyclones and anti-cyclones, or provide a measure of the number of storm systems or their intensity.

The Lagrangian approach or feature tracking dates back to the mid-nineteenth century when scientists started to classify and plot individual storms using synoptic weather maps (Mohn, 1870; Bergeron, 1950). Several manual studies have been performed using synoptic charts (Pettersen, 1956; Klein, 1957; Whittaker and Horn, 1982), but this is a very time consuming process. The onset of powerful computer modelling allowed a more global, synoptic approach based on statistical methods through the so-called ‘synoptic objective analysis’ (Murray and Simmonds, 1991; Serreze et al., 1993; Sinclair, 1994; Hodges, 1995; Hoskins and Hodges, 2002; Zhang et al., 2004; Inatsu, 2009). This type of analysis made it possible to employ algorithms that identify storms as local minima or maxima of parameters, such as sea level pressure

or relative vorticity and link them together to form storm trajectories. Several methods have been developed and differences can occur for the diagnostics produced by different methods.<sup>8</sup> However, this is more often than not associated with the use of different fields, e.g. MSLP, geostrophic or relative vorticity, data on different pressure levels or the choice of which storms go into the statistics. A good review of storm tracks from the Lagrangian perspective for both the current and future climates can be found in Ulbrich et al. (2009).

### 3.2. Why do storms matter?

Extratropical storms impact a sea-ice region by bringing heat/moisture and fostering sea-ice melting. The sea-ice changes may in turn affect the downstream development of storm tracks (Honda et al., 1999; Mesquita et al., 2010a). This two-way interaction process alters local albedo and fluxes of heat and temperature, with large implications for the Northern Hemisphere (Yamamoto et al., 2006). This interaction may also affect the large-scale variability through a Rossby wave response (Alexander et al., 2004).

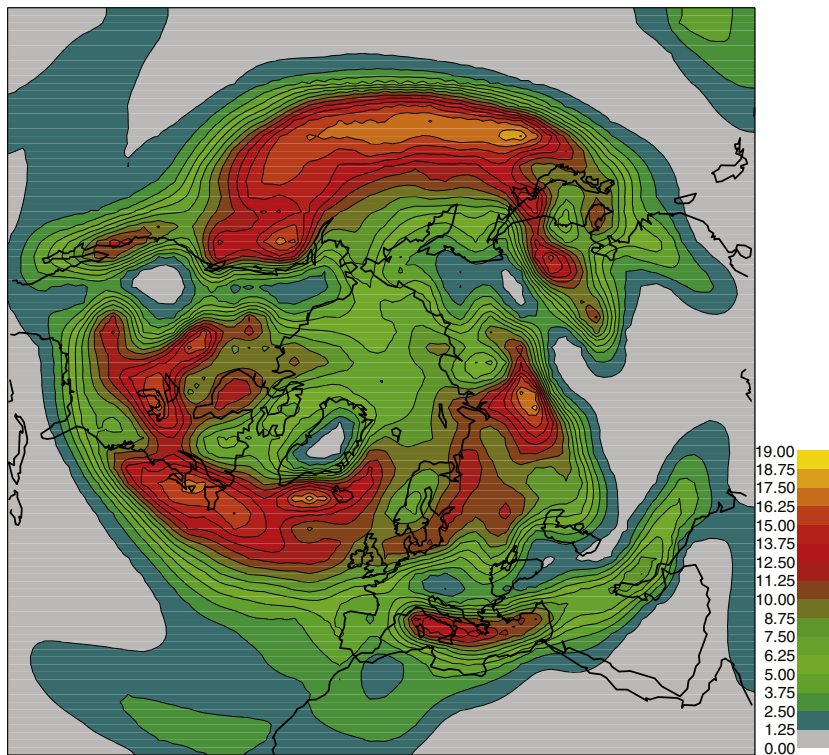
### 3.3. Storm observations

As atmospheric observations are inhomogeneous in space and time, even in the current modern satellite period, the best way to obtain detailed information of the distribution and properties of storms is from atmospheric re-analyses. These combine modern Numerical Weather Prediction (NWP) models and historical observations in a dynamically consistent way using data assimilation. Boreal winter, December–February (DJF), storm track density based on the NCEP–CFRS (Saha et al., 2010) reanalysis (Fig. 5) shows two major zones of storm density maxima over the North Atlantic/eastern North America and the North Pacific that extend into high latitudes (see also Hoskins and Hodges, 2002; Mesquita et al., 2008; Ulbrich et al., 2009; Gulev et al., 2001). In general there is good agreement among reanalysis on NH. This is partly because they are well constrained by available observations, although this becomes less so in early periods (Hodges et al., 2003, in press; Wang et al., 2006; Bromwich et al., 2007).

An important question is whether the number and/or intensity of Northern Hemisphere extratropical cyclones have changed over the past decades. There is no general agreement on this. Extratropical storm studies have shown a decrease in the number of cyclones in the North Atlantic (e.g.: Gulev et al., 2001; Raible et al., 2008; in mid-latitudes only, McCabe et al., 2001; Wang et al., 2006) and in the North Pacific (e.g.: Gulev et al., 2001; in mid-latitudes only McCabe et al., 2001). In contrast, Graham and Diaz (2001) show an increase in the cyclone frequency in the North Pacific, Simmonds et al. (2008) found positive trends in the number of strong systems in the Arctic and Zhang et al. (2004) found an increase in the number of cyclones entering the Arctic – in agreement with

<sup>8</sup> Two noteworthy comparison projects are: EU-COST Action 733 Harmonization and Applications of Weather Type Classifications for European Regions; and IMILAST – Intercomparison of Mid Latitude Storm Diagnostics. More information and further references can be found on their websites.





**Fig. 5.** Northern Hemisphere winter (DJF) track density computed from the 850 hPa relative vorticity from the NCEP–CFSRR re-analysis for 1979–2009. Densities are number density per month per unit area, where the unit area is equivalent to a 5 degree spherical cap ( $\approx 1 \times 10^6 \text{ km}^2$ ). Methodology same as Hoskins and Hodges (2002).

Sorteberg and Walsh (2008) who found positive trends in cyclone activity for cyclones entering the Arctic in three out of four seasons. Other studies find an increase in the frequency of extreme cyclones in the North Pacific and North Atlantic (Geng and Sugi, 2001; Paciorek et al., 2002). Some regional studies do not show statistically significant trends (Mesquita et al., 2010b; Simmonds and Keay, 2009; Wang et al., 2006). In fact, Simmonds and Keay (2009) show that the strength rather than the number of cyclones in the Arctic basin plays a role in the observed changes there.

The differences in trends among these studies depend on the region selected, and on the tracking schemes and reanalysis data used (Raible et al., 2008; Ulbrich et al., 2009). However, the largest difference is likely due to the use of different fields – MSLP, geostrophic vorticity or relative vorticity – and the spatial scales identified. For example, if storm intensity is measured as the pressure minima, trends in the intensity could partly reflect large-scale circulation changes. The MSLP is strongly influenced by large spatial scales, such as the Icelandic low, and strong background flows (Sinclair, 1994). Thus the identification of features in unfiltered MSLP tends to be dominated by large-scale features and biased toward the slower moving systems. Also in part, the uncertainty in the trends may be related to artificial shifts in reanalysis caused by major changes in the observing systems, such as the introduction of satellite data (Bengtsson et al., 2004; Bromwich et al., 2007).

It is interesting to note that Harnik and Chang (2003) suggests that the intensification of storms in the North

Atlantic and North Pacific found in reanalysis data is weaker than that of sonde data. In the Pacific storm track entrance and exit regions, they show that there are no significant positive trends. They add that sonde data “...show a positive trend over Canada, consistent with a Pacific storm track intensification and northeastward shift, but lack of data over the storm track peak prevents drawing any strong conclusions.” Re-analyses, particularly the older lower resolution ones, may well underestimate the storm intensities, in particular the low level winds (Hodges et al., in press).

From an Eulerian perspective, the IPCC Fourth Assessment Report mentions that several results suggest an increase in the cyclone activity in the Northern Hemisphere. These changes have been found in NCEP-reanalysis-based eddy statistics for the North Pacific (Nakamura et al., 2002; Chang, 2003); eddy meridional velocity variance at 300 hPa and other variables (Chang and Fu, 2002; Paciorek et al., 2002). Despite an increase in the amount of eddy kinetic energy due to increased efficiency in the conversion from potential to kinetic energy (Hu and Wu, 2004) and enhanced MSLP variance over the Pacific (Graham and Diaz, 2001), there are significant uncertainties with such analyses. Some studies suggest that storm track activity during the last part of the 20th century may not be more intense than the activity before the 1950s (Bromirski et al., 2003; Chang and Fu, 2003). Eddy meridional velocity variance at 300 hPa in the NCEP Reanalysis seems to be biased low before the mid-1970s, especially over east Asia and the western USA (Harnik and Chang, 2003). The increase in eddy variance in the NCEP reanalysis

data are nearly twice as large as that computed from radiosonde observations, with better agreement for the Atlantic storm track exit region. This could well be due to the introduction of satellite observations in the late 1970s (Bengtsson et al., 2004).

The IPCC report also mentions that major differences between the NCEP reanalysis temperature variance at 500 hPa over Asia and radiosonde data also cast doubts on the magnitude of the increase in storm track activity over the Pacific (Iskenderian and Rosen, 2000; Paciorek et al., 2002). In addition to that, station pressure data over the Atlantic–European sector show a decrease in storminess from high levels during the late 19th century to a minimum around 1960 and then a rapid increase to a maximum around 1990, followed again by a slight decline (Alexandersson et al., 2000; Barring and von Storch, 2004). However, the noise present in the observations makes the detection of long-term changes in extratropical storm activity difficult. The IPCC report refers to the analysis of regional storminess in relation to spatial shifts and strength changes in teleconnection patterns, as a more relevant approach.

### 3.4. Storm projections

#### 3.4.1. Lessons learned

Recent studies indicate that climate change may cause NH storm tracks to shift several degrees poleward (Geng and Sugi, 2003; Fischer-Bruns et al., 2005; Yin, 2005; Bengtsson et al., 2006; Rinke and Dethloff, 2008; Schuenemann and Cassano, 2010). This is in accordance with previous results (Schubert et al., 1998). Some coarse-resolution studies, such as Lambert and Fyfe (2006), shown a possible reduction in NH mid-latitude storms, but with a decrease in central pressures in the storm systems. Lambert and Fyfe (2006) find no poleward shift of the storm tracks, but this was not a Lagrangian study and only counted MSLP minima. Some Lagrangian studies also show a reduction in cyclone numbers in the future likely associated with the reduced baroclinicity. However, there is no apparent increase in cyclone intensities (Bengtsson et al., 2006, 2009; Catto et al., *in press*) in terms of winds even though there is an increase in latent heat release from larger precipitation. However, larger changes can occur for particular regions, though with greater uncertainty: for example Inatsu and Kimoto (2005) show a more active storm track in the western Pacific in the future but weaker elsewhere, while Fischer-Bruns et al. (2005) find the opposite.

In summary, the most consistent results from the majority of the current generation of models are a poleward shift of NH storm tracks in a future warmer climate. However, what have we actually learned when it comes to storm projections? The answer to this question remains unclear. Some of the inconsistencies in the different papers could be related to the variable used for tracking the storms; the model resolution; and in the case of trends, the region selected for averaging. Also, the use of different models which have different degrees of polar surface and upper tropical troposphere warming together with different levels of change of the MOC can produce different storm track responses. An approach and framework for which robust conclusions can be drawn on cyclone changes, in cases where warming proceeds

at a different rate in different regions, is given in the semi-idealised experiments of Lim and Simmonds (2009). The authors show that, for the Southern Hemisphere, the warming of the tropical upper troposphere increases static stability and thus reduces baroclinicity in low and midlatitudes. This would help explain the decrease of extra-tropical cyclones in mid-latitudes and the slight increase in higher latitudes could be explained by the increased meridional temperature gradient. Although most relevant for the SH, the recovery of stratospheric ozone, projected for the middle of this century, introduces another uncertainty. It is included in some models and not in others and has an impact on the upper tropospheric temperature gradient and baroclinicity and thus impacts the poleward shift of storm tracks (Son et al., 2008).

Below we summarise a few studies related to sea-ice changes and storm tracks:

- Knippertz et al. (2000) found a pronounced north and eastward shift over Europe and the north-east Atlantic based on a 240 year run of the ECHAM4/OPYC3 coupled ocean–atmosphere model, at a spectral resolution of T42 (19 vertical levels) using the transient greenhouse gas forcing in the IPCC IS92a scenario. The storm tracking was based on the storm-identification scheme of Haak and Ulbrich (1996) and the variable used was the 1000 hPa geopotential height. They also found a decrease in the number of weak cyclones and an increase in deep cyclones. The deep cyclones over the whole Atlantic increases by 40% in winter compared to the control period. They hypothesised that the cyclone change signal is due to changes in upper-tropospheric baroclinicity. Strong wind speeds found over the Hudson Bay and the Greenland Sea (in the boundary layer) are connected to the reduction in winter mean sea-ice cover, thus leading to a reduction of static stability and over Greenland, a reduction in surface roughness. The decrease in static stability is connected with the enhanced sensible heat flux from the ice-free surface and the turbulent transport of momentum from the free atmosphere downward into the planetary boundary layer – producing high wind speeds. The increase of extreme winds over Hudson Bay can be accounted by the higher frequency of weak local cyclones there. In general, the authors find larger wind extremes in the areas where sea ice has retreated.
- Geng and Sugi (2003) used a higher-resolution (about 100 km resolution) atmospheric GCM (AGCM) with time-slice experiments and find a decrease in cyclone density in the mid-latitudes in a warmer future climate (DJF season), associated with baroclinicity changes in the lower troposphere. In the control run, the atmosphere was forced by the observed SST and sea ice of 1979–98 and present-day CO<sub>2</sub> and sulphate aerosol concentrations. In the global warming run, the atmosphere is forced by the observed SST and sea ice of 1979–98 plus the monthly mean anomalies of SST and sea ice at about the year 2050 obtained from a transient climate change experiment with the GFDL<sup>9</sup> coupled ocean–atmosphere model with a resolution of R15. They also find that the density of strong

<sup>9</sup> Geophysical Fluid Dynamics Laboratory.

- cyclones increases while the density of weak and medium-strength cyclones decreases, intensities based on MSLP.
- [Leckebusch and Ulbrich \(2004\)](#) found a slight reduction of tracks compared to the present day climate (6% reduction) in their study of the A2a and B2a scenarios. They also found a reduction of 19% over the Denmark Strait, Iceland and the Norwegian Sea. The B2a scenario shows similar patterns except for increased track density between 55 and 60 N – the region where there was a small decrease in the A2a scenario. Extreme storms are shifted southwards and less pronounced in the B2a scenario. Extreme cyclones show an increase of 50% south of Iceland for the A2a scenario. The A2a scenario shows a clear signal towards more intense cyclones affecting Western and Central Europe compared to the present day climate conditions. In contrast, [Knippertz et al. \(2000\)](#) shows increased cyclone frequency above Northern Europe. [Schubert et al. \(1998\)](#) also identified a shift of the cyclone track density northeastward. [Leckebusch and Ulbrich \(2004\)](#) explain that these differences between HadCM and ECHAM runs are not clear and could be related to different transient fluxes of energy due to altered baroclinicity changes in the different models.
  - [Yin \(2005\)](#), using coupled simulations of the current and future climate (A1B scenario) from the IPCC AR4, found a poleward and upward shift and intensification of storm tracks. Storm tracks were analysed using eddy kinetic energy (Eulerian approach), filtered to retain variability on synoptic time scales of 2–8 days. All but four models showed a poleward shift in NH winter: ECHAM5/MPI-OM, MRI-CGCM2.3.2, GISS-AOM and INM-CM3.0. The poleward and upward shift of storm tracks can be associated with the poleward and upward expansion of baroclinic regions, although the surface temperature gradient is reduced. This can be related to the changes in the low and upper level temperature gradients associated with the Arctic warming from the sea-ice albedo feedback and the tropical upper tropospheric warming respectively. In the North Atlantic this is further complicated by the slowing down of the Meridional Overturning Circulation (MOC) which results in a reduced warming in the North Atlantic which can enhance the local baroclinicity ([Brayshaw et al., 2009](#)). Different coupled models show different degrees of MOC slowdown.
  - [Bengtsson et al. \(2006\)](#) investigated the storm track changes in the current and a future scenario (A1B) using the ECHAM5 coupled climate model. Differently from previous studies, they find indications of a poleward shift in the storm tracks in the Northern Hemisphere winter and a strengthening of the storm track north of the British Isles ([Fig. 8](#)). They find a minor reduction in the number of weaker storms and no indication of more intense storms. One of the main differences in their study is the choice of the tracking variable: relative vorticity instead of the mostly used MSLP. The increase in storm intensity found in previous studies may reflect changes in the large-scale MSLP background ([Bengtsson et al., 2006](#)). When using the MSLP a better approach is to use the geostrophic vorticity (Laplacian of the MSLP) ([Sinclair, 1994](#); [Simmonds and Keay, 2009](#)) or filtered MSLP fields although this will depend on the form of the filter.
  - [Bengtsson et al. \(2009\)](#) investigate extratropical cyclones and how they may change in a warmer climate in detail with a high-resolution version of the ECHAM5 global climate model. A spectral resolution of T213 (63 km) is used for two 32-year periods at the end of the twentieth and twenty-first centuries and integrated for the A1B scenario in time slice mode using SSTs from the lower resolution coupled simulations. The authors find that, for the twenty-first century, changes in the distribution of storms are very similar to previous studies. There is a small reduction in the number of cyclones but no significant changes in the extremes of wind and vorticity in both hemispheres. There are larger regional changes in agreement with previous studies. The largest changes are in the total precipitation with a significant increase in the future. Cumulative precipitation along the tracks of the cyclones increases by around 11% per track – that is twice the increase in global precipitation. The increase in the most extreme precipitation is close to the globally averaged increase in column water vapour (about 27%).
  - [McDonald \(in press\)](#) also found a poleward shift in the storm tracks in some seasons and regions and fewer cyclones in winter and spring. Also found was an increased frequency of strong winds over the British Isles associated with a southward shift of the northeast end of the North Atlantic storm track. The shift seems to be related to an increase in baroclinicity and a southward shift of the jet.
- Changes of cyclone activity are closely related to the change of baroclinicity in the lower troposphere, which are mainly related to the changes in horizontal and vertical temperature distributions (stability) in the atmosphere under global warming ([Geng and Sugi, 2003](#)). [Geng and Sugi \(2003\)](#) find that in the Northern Hemispheric mid-latitudes, the decrease of baroclinicity is mainly caused by the decrease of the meridional temperature gradient. In the Atlantic Arctic sector there is also a significant impact on the baroclinicity from the enhanced static stability (see [Zahn and von Storch, 2010](#) later in the text). However, [Geng and Sugi \(2003\)](#) point out that other factors may also play a role in cyclone activity changes. The increase in the atmospheric moisture due to global warming may be an important factor that influences the changes in cyclone activities, as pointed out by [Hall et al. \(1994\)](#); [Lambert \(1995\)](#); [Sinclair and Watterson \(1999\)](#), and [Carnell et al. \(1996\)](#). However, although increased moisture (acting alone) might increase the cyclone activity, especially the intensity of extreme storms, there are other processes that can affect the cyclone activity. In fact changes in cyclone activity and intensities in a warmer climate will be the consequence of changes in competing process such as the reduction (increase) in baroclinicity in the lower (upper) troposphere and the increase in latent heat release ([Sinclair and Watterson, 1999](#); [Bengtsson et al 2009](#)) and in the north Atlantic the slowing MOC. The changes may also be resolution dependent and may of the low resolution studies will not represent cyclones correctly. Also, [Catto et al. \(in press\)](#) using 2 and 4 times CO<sub>2</sub> simulations with the HiGEM model actually find a decrease in intensities in the NH winter with the N. Atlantic showing contradictory results between 2 and 4 times CO<sub>2</sub> related to the relative changes on the upper and lower tropospheric

temperature gradients. Bengtsson et al. (2009), using T213 simulations with the ECHAM5 model of the current and future climates under the A1B scenario, find no overall increase in the intensities of storms in the NH winter either in terms of winds or vorticity even though the precipitation intensity increases significantly. Rather, there are significant changes on a region by region basis. The result of this review indicates that the frequency of extratropical cyclones is likely to decrease due to global warming, while there is still considerable uncertainty in the likely changes in the storm intensities. We note that there are considerable differences among the studies concerning the possible future behaviour of extratropical cyclones, in particular at the regional scale.

#### 3.4.2. Polar lows

Most previous studies of cyclones in the extra-tropics and high latitudes have focused on synoptic scale cyclones, their trends and changes in the future with increasing temperatures. The issue of meso-cyclones, namely polar lows has received less attention apart from some limited observational studies (Businger, 1985; Harold et al., 1999). The resolution of current climate models is insufficient to simulate all but the largest scale polar lows. A limited number of studies have downscaled re-analysis and climate models to resolutions more capable of simulating polar lows, albeit not their intensities (Zahn et al., 2008; Zahn and von Storch, 2010). These have shown no significant trends of polar lows in the North Atlantic high latitudes over the period of the NCEP–NCAR re-analysis. However, it is unclear if using a different re-analysis would alter significantly results bearing in mind the uncertainties between the older re-analysis in the polar regions, particularly in their depiction of clouds and their associated radiation impacts (Bromwich et al., 2007). From the downscaling of the ECHAM5/MPI-OM AR4 integrations Zahn and von Storch (2010) found that there is projected to be a decrease in polar lows in the future associated with reduced low level thermal gradients with the retreat of sea ice and an increase in the static stability. It is unclear how this picture may change with a different model but this may become more clear with the use of new higher resolution re-analysis, such as the NCEP–CFSR (Saha et al., 2010), and the gradual increase in climate model resolutions.

## 4. Sea ice and storms: observations and idealised experiments

The study of Deser et al. (2000) explicitly notes on page 617 that:

*The temporal and spatial relationships between the SLP and ice anomaly fields are consistent with the notion that atmospheric circulation anomalies force the sea ice variations. However, there appears to be a local response of the atmospheric circulation to the changing sea ice cover east of Greenland. Specifically, cyclone frequencies have increased and mean SLPs have decreased over the retracted ice margin in the Greenland Sea, and these changes differ from those associated directly with the North Atlantic oscillation.*

To zero order, the atmosphere forces sea ice (see also e.g. Partington et al., 2003; Koenig et al., 2006), but the sea ice

feeds back onto the atmosphere (Honda et al., 1999; Alexander et al., 2004; Magnusdottir et al., 2004). Here, we will focus on the sea-ice impact on the atmosphere, specifically on storms in this section.

### 4.1. Sea ice and storm observations

Simmonds and Keay (2009) studied the relationship between the September Arctic sea-ice reductions and storm track changes. Using an algorithm based on the Laplacian of the MSLP (Murray and Simmonds, 1995), they showed an increase in strength, rather than the number, of observed cyclones in the Arctic basin. Hence, stronger cyclones could provide more mechanical forcing on sea ice in the Arctic – which could result in sea ice being dispersed and reducing even more during September.

Overland and Pease (1982) using an observational approach show a decrease in the number of cyclones with latitude in all months and division into two storm tracks, one propagating north–northeast and another entering the southern Bering Sea. Composite cyclone charts for the five heaviest and five lightest ice winters show a shift in cyclone centres toward the west in light ice years. The relation of sea ice extent and the location of cyclone tracks is consistent with previous observations that advance of the ice edge in the Bering Sea is dominated by wind-driven advection and that southerly winds associated with cyclone tracks to the west inhibit this advance.

Deser et al. (2000) is a pioneering study relating observational changes in sea-ice conditions to changes in the atmosphere, including storm tracks. By using data from an automated storm tracking algorithm and the MSLP field for tracking (Serreze, 1995; Serreze et al., 1997), they showed that in low sea-ice periods there is a poleward shift of storm tracks in the North Atlantic. The shift is more located northwestward into the Greenland Sea – the region of the climatological mean Icelandic low and east of Greenland where the sea ice concentrations have decreased (Deser et al., 2000). The authors also hypothesise that the increase in the number of storms over the reduced sea-ice area in the Greenland Sea is the result of the variations of surface energy flux into the atmosphere from the lower boundary.

### 4.2. Sea ice and storms: idealised experiments

The impact of sea-ice changes on storm tracks, storminess and atmospheric variability patterns is generally analysed using atmospheric general circulations models (AGCMs) forced by observed, idealised or projected future sea-ice changes (Marshall et al., 2004; Magnusdottir et al., 2004; Seierstad and Bader, 2009; Deser et al., 2010). Compared to coupled ocean–atmosphere–sea-ice models this set-up of experiments is idealised, because no impact of the atmosphere on the sea-ice concentration (SIC) is considered. The SICs are prescribed every month. In addition to that, other forcing like greenhouse gas concentrations and SSTs are often kept constant. The advantage of this kind of set-up is that the exclusive impact of certain sea-ice changes on atmospheric fields can be analysed. Normally, at least two experiments are carried out with different sea-ice concentrations in a region. The experiments are run in ensemble mode: the experiments are repeated several times with the same boundary condition

(same monthly sea-ice concentration), but with slightly different initial conditions in the atmosphere. The ensemble mode technique helps to separate the signal (sea-ice change) from the noise in the system. The two means of each ensemble can then be compared and analysed.

An example of sea-ice sensitivity experiments is shown in Figs. 9 and 10. The ECHAM5 AGCM has been run at T213 (63 km) resolution in time slice mode for two 30 year periods for the current climate, one with sea ice and one without. The sea ice concentration and SST are taken from a coupled simulation of the ECHAM5 model for the current climate. The simulation with sea ice is the one used in Bengtsson et al. (2009). All other forcing are the same. Removing sea ice causes significant decreases in the track density in the Arctic region. Also Seierstad and Bader (2009) find a reduction in storminess towards the Arctic when forcing the AGCM ECHAM5 with a projected future Arctic sea-ice reduction and no changes in greenhouse gases. Zahn and von Storch (2010) find a decreased frequency of polar lows in the Arctic north Atlantic in global warming simulations. This indicates that a reduction in Arctic sea ice at least contributes to the reduced Arctic storminess projected by future simulations. Another feature of the T213 sea-ice experiments is the northward shift of the track density pattern in the Pacific and a southward in the Atlantic basins. Fig. 10 shows the mean cyclone speed difference between these two experiments. It shows a similar structure as that of Fig. 9 for the track density.

Most publications dealing with idealised experiments show that Arctic sea-ice changes have an impact on storminess (i.e.: location and intensity). Kvamstø et al. (2004) used the band-pass-filtered 500 hPa geopotential height (2–10 day frequency) to study storms in an AGCM idealised simulation. Reduction in Labrador sea-ice was associated with a northward shift in the North Atlantic storm tracks. Magnusdottir et al. (2004) carried out experiments using the NCAR AGCM with specified sea-ice anomalies confined to the North Atlantic sector. They exaggerated the observed trends in sea-ice in the Labrador and Greenland seas. They find a weaker, southward-shifted and more zonal storm track. Alexander et al. (2004) study the influence of realistic Arctic sea-ice anomalies in the atmosphere during winter using the atmospheric CCM 3.6 model. They find a weaker North Atlantic storm track due to reduced (enhanced) ice cover to the east (west) of Greenland. Seierstad and Bader (2009) forced the AGCM ECHAM5 with a projected future Arctic sea-ice reduction and no changes in greenhouse gases. They find significant reductions in storminess during winter in both midlatitudes and towards the Arctic, especially pronounced during March.

Murray and Simmonds (1995) show, in idealised sea-ice experiments with a low horizontal resolution of R21 and nine vertical levels, that decreases in sea-ice concentration produced a monotonic but nonlinear warming in the lower troposphere and weakening and southward contraction of the mid-latitude westerlies. There was a significant decrease in the speeds and intensities of cyclonic systems north of 45 N but little overall change in areal densities or in the arrangement of the major storm tracks. At 500 hPa the effect of the high-latitude warming (and low latitude-cooling) was to reduce the depth of the polar vortex and reduce the strength of the westerlies north of about 45 N. No significant

expansion of cyclonic activity occurred in the central Arctic. The major storm tracks remained evident through the series of experiments, but quantitative measures of cyclone behaviour displayed considerable variation with open water fraction. A reduction of the speeds and intensities of storms was evident north of 55 N at most longitudes and north of about 45 N in their zonal averages. The reduction in cyclone speeds is considered to be a direct effect of changes in the steering flow, while the reduction in intensities would be an indirect effect of reduced mid-tropospheric wind speeds brought about via decreased baroclinicity. A significant finding of this work is the relative insensitivity of the storm track pattern and of overall cyclone numbers to changes in sea ice concentration, but the model resolution may be too low to properly simulate the cyclones.

Geng and Sugi (2003) considered a future run using SST and sea-ice anomalies at about the year 2050 contrasted against a control run forced with SST and sea-ice from 1979 to 98. An objective cyclone identification was used (Geng and Sugi, 2001). A decrease of 7% in the number of cyclones was observed in the wintertime (DJF), whereas a change of 3% for the summertime (JJA). The density of strong cyclones increases by more than 20% in the summer over the eastern coasts of Asia and North America. These changes are due to the decrease of meridional temperature gradient. The decrease in the Atlantic is also accompanied by an apparent shift of the storm tracks southwards in the winter. Our Fig. 9 shows the track-density change for a run where the sea-ice has been removed. There is a clear southward storm track shift apparent in the North Atlantic. By comparing with the results of Geng and Sugi (2003) this gives a hint that at least some part of the storm track shift is induced by the removal of the Arctic sea-ice cap, possibly caused by the decrease of the meridional temperature gradient. Strong cyclones shift northward in the northern North Pacific and shift southeastward in the northern North Atlantic region. This is in contrast with Knippertz et al. (2000) who found increasing frequencies of strong cyclones but with a northward shift of the strong cyclone activity in the North Atlantic.

To summarise we have seen that there are some indications that there is a northward shift of the NH storm-track maxima in mid-latitudes in global warming simulations. The idealised sea ice experiments indicate that at least part of the Pacific mid-latitude poleward storm-track shift is caused by the retreat of the Arctic sea-ice. The situation is less clear for the Atlantic basin. Arctic sea-ice reductions may locally reduce storminess.

## 5. The North Atlantic Oscillation

### 5.1. Teleconnection patterns: a brief introduction

The term teleconnection is often used in atmospheric sciences to describe the climate links between geographically separated points (Nigam, 2002), forming large-scale atmospheric circulation patterns. Common techniques to define teleconnections are, for example, “Contemporaneous correlations between sea level pressure or geopotential heights on a given pressure surface at widely separated points on earth” (Wallace and Gutzler, 1981) or Empirical Orthogonal

Functions (EOFs) (Preisendorfer, 1988) analysis of SLP or geopotential height. It has to be mentioned that caution should be used when trying to interpret these statistically derived EOF modes (see Dommenges and Latif, 2002; Dommenges, 2007). One of the main differences between the two methods is that the EOF method focuses on regions that account for a substantial portion of the temporal variance rather than just those which exhibit strong correlations with distant points (Nigam, 2002).

Of the many teleconnection patterns,<sup>10</sup> the North Atlantic Oscillation has gained the most attention in current literature, when it comes to the impact of Arctic sea-ice changes on teleconnection patterns. Although this review focuses on the North Atlantic Oscillation (NAO), it is important to bear in mind that other NH teleconnection patterns (e.g. the Pacific/North American (PNA) pattern) may also be affected by changes in Arctic sea ice – however, to our knowledge, most papers describing the PNA discuss the impact of this teleconnection pattern on the Arctic sea ice (see e.g. L'Heureux et al., 2008), rather than the opposite. Arctic sea-ice changes may also have an impact on the atmosphere in the Pacific region by exciting a large-scale Rossby wave train (Honda et al., 1999; Budikova, 2009), but the response projects only on the NAO (Mesquita et al., 2010a).

## 5.2. NAO and its associated meteorological phenomena

The North Atlantic Oscillation (NAO) is a large-scale alternation of atmospheric mass with centres of action near the Icelandic low and the Azores high (Hurrell, 2002). It characterises the negative correlation between these two locations (Barry and Carleton, 2001). The NAO index has been defined as the normalised mean sea level pressure difference between Ponta Delgada, Lisbon or Gibraltar and Stykkisholmur/Reykjavik on a monthly or seasonal resolution or as the principal component time series of the leading EOF of monthly to seasonal (December through March) sea level pressure anomalies over the Atlantic sector (Fig. 6) (Hurrell, 1995a; Jones et al., 1997). A positive NAO phase is associated with an anomalous deep Icelandic low and intense subtropical high.

The NAO is the dominant mode of atmospheric variability in the North Atlantic sector throughout the year (Barnston and Livezey, 1987), although it is most pronounced during winter and accounts for more than one-third of the interannual winter total variance in sea level pressure (Cayan, 1992; Hurrell et al., 2003, and references therein). The NAO is associated with various alterations of meteorological parameters and with changes in the Atlantic ocean and ecosystems (Hurrell and van Loon, 1997; Alexandersson et al., 1998; Hurrell, 2002; Hurrell and Deser, 2009; Frankcombe et al., 2010; Woollings, 2010). Some of these are as follows:

- Biology; investigations have established links between the NAO and the biology of the North Atlantic. For example, the change in biomass and species composition of plankton and zooplankton and the biomass, distribution and growth of fish species (Drinkwater et al., 2003).

- Temperature; the NAO exerts a dominant impact on winter temperatures across much of the Northern Hemisphere (Hurrell, 2002). While the NAO index is positive, enhanced westerly flow in the Atlantic sector during winter brings relatively warm, moist, maritime air over the northern part of Europe and large parts of Asia. At the same time stronger northerlies over Greenland and over the north-eastern part of Canada bring cold air southward over the western Atlantic sector. The NAO explains about one third of the interannual winter temperature variability in the NH (Hurrell, 2002).
- SST; fluctuations in the NAO are associated with large-scale SST anomalies over the North Atlantic, with a tripole-like pattern: cold anomalies in the subpolar North Atlantic, warm anomalies in the centre and cold anomalies south of  $\approx 30$  N (see e.g. Rodwell et al., 1999; Hurrell, 2002; Visbeck et al., 2003).
- Sea ice; during the positive phase of the NAO the Labrador sea-ice extent is further south due to enhanced northerly winds in the Labrador Sea region, while the Greenland Sea ice extent is pushed further north because of stronger southerly winds in this region.
- Storm tracks; the NAO is intimately linked to changes in the extratropical storm tracks and jet stream. When the NAO index is positive, a northerly shift in the mean winds and storm track regime is observed: more intense and frequent storms near Iceland and the Norwegian Sea (Serreze et al., 1997; Deser et al., 2000) lead to wetter conditions over Scandinavia and northern Europe. The reduced storms over the Mediterranean and northern Africa lead to drier conditions there. Fig. 5 shows the difference in track density<sup>11</sup> for the NAO for the NH winter computed from the 850 hPa relative vorticity from the NCEP–CFSR re-analysis for 1979–2009 and based on the CPC NAO index.<sup>12</sup>

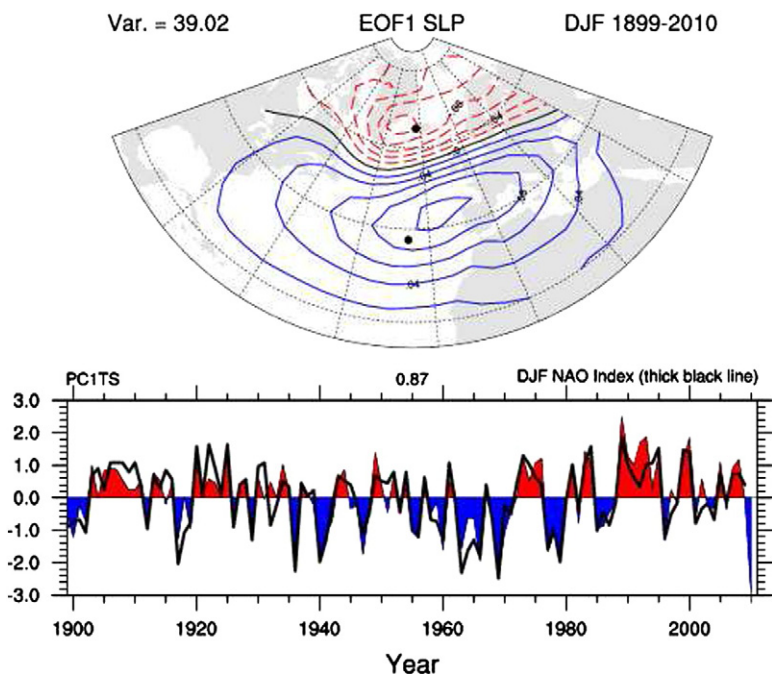
## 5.3. NAO in observations

The NAO does not seem to vary on any preferred time scale. Large interannual changes can occur from one winter to the next (Hurrell et al., 2003). There are, however, periods when anomalous NAO-like circulation patterns persist over quite a few consecutive winters. There was a strong negative trend from the late-1940s to the mid-1960s and an increasing trend that started in the mid-1960s till the mid-1990s (see e.g. Stephenson et al., 2000). The spectrum of the winter-mean NAO index is slightly “red”, with power increasing with period, but does not show significant peaks (Hurrell et al., 2003). Feldstein (2000) points out that the interannual variability of the NAO can be interpreted as being a stochastic process. Whereas the trend and increase in the variance of the NAO index from 1968 through 1997 was greater than would be expected from internal atmospheric variability alone, while its behaviour during the first 60 years of the 20th century was consistent with atmospheric internal variability (Feldstein, 2002). Using coupled atmosphere–ocean models

<sup>10</sup> See Wallace and Gutzler (1981) for a list of different teleconnection patterns.

<sup>11</sup> Densities are number density per month per unit area, where the unit area is equivalent to a 5 degree spherical cap ( $\approx 1 \times 10^6 \text{ km}^2$ ).

<sup>12</sup> See <ftp://ftp.cpc.ncep.noaa.gov/wd52dg/data/indices/teleindex.nh>.



**Fig. 6.** The North Atlantic Oscillation pattern for winter (DJF) from 1899 to 2010. Above: the first EOF; and below: the principal component time series of the leading EOF of winter SLP anomalies over the Atlantic sector (20–80 N, 90 W–40E). The station based index is given by the thick black line.

Source: Figure downloaded from the NCAR Climate Analysis Section on their website at <http://www.cgd.ucar.edu/cas/jhurrell/indices.info.html#naopcdfj>.

Semenov et al. (2008) show that the recent observed increasing NAO trend till 1995 cannot be statistically distinguished from natural internal variability generated within the coupled atmosphere–ocean system.

#### 5.4. What are the mechanisms which govern the NAO variability?

There is no consensus on the processes that are responsible for the observed variations in the NAO. There is evidence that much of the atmospheric circulation variability in the form of the NAO arises from internal atmospheric processes. Atmospheric general circulation models (AGCMs) forced by climatological sea surface temperature, and fixed atmospheric trace-gas concentrations, display NAO-like fluctuations (Saravanan, 1998). The governing dynamical mechanisms are eddy mean flow interaction at the exit region of the Atlantic storm track and eddy–eddy interaction between baroclinic transients and low-frequency variability (Hurrell, 1995b). The barotropic component of the zonal wind anomalies are driven by the convergence of eddy momentum fluxes (Thompson et al., 2003). On the other hand, it has long been recognised that fluctuations in SST and the strength of the NAO are related (Bjerknes, 1964), and there are clear indications that the North Atlantic Ocean varies significantly with the overlying atmosphere.

The leading mode of SST variability over the North Atlantic during winter consists of a tri-polar pattern with a cold anomaly in the subpolar region, a warm anomaly in the middle latitudes centred off of Cape Hatteras, and a cold subtropical anomaly between the equator and 30 N (Deser and Blackmon, 1993; Rodwell et al., 1999). The strength of the

correlation increases when the NAO index leads the SST, which indicates that SST is responding to atmospheric forcing on monthly time scales (Battisti et al., 1995). But SST observations also display interannual and decadal responses (Sutton and Allen, 1997; Visbeck et al., 1998), which may indicate that decadal/multi-decadal variations of the ocean surface have an impact on the atmosphere. Rodwell et al. (1999) forced their AGCM by the observed, global SSTs and sea ice concentration over the past 50 years (see also Latif et al., 2000; Mehta et al., 2000). They captured much of the multi-annual to multi-decadal variability in the observed NAO index since 1947, including about 50% of the observed strong upward trend over the past 30 years. To confirm that North Atlantic SSTs force the NAO they drove their AGCM with a North Atlantic SST pattern similar to the observed SST tripole. They found that the atmospheric response in mean sea level pressure between model simulations with positive and negative versions of the SST tripole is similar to the classical NAO pattern. They concluded that much of the multi-annual to multi-decadal variability of the winter NAO over the past half century may be reconstructed from a knowledge of North Atlantic sea surface temperatures.

More recent studies conclude that NAO variability is closely tied to SSTs over the tropical ocean (Hoerling et al., 2001; Bader and Latif, 2003; Hoerling et al., 2004; Hurrell et al., 2004; Bader and Latif, 2005). Hoerling et al. (2001) presented evidence that North Atlantic Climate change since 1950s is linked to a progressive warming of tropical SSTs. They argue that the ocean changes alter the pattern and magnitude of tropical rainfall and atmospheric heating, the atmospheric response to which includes the spatial structure of the NAO.

Watanabe and Nitta (1999) have suggested that land processes are responsible for decadal changes in the NAO. They find that the change toward a more positive wintertime NAO index in 1989 was accompanied by large changes in snow cover over Eurasia and North America. Moreover, the relationship between snow cover and the NAO was even more coherent when the preceding fall snow cover was analysed, suggesting that the atmosphere may have been forced by surface conditions over the upstream land mass. Watanabe and Nitta (1998) reproduce a considerable part of the atmospheric circulation changes by prescribing the observed snow cover anomalies in an AGCM. Bojariu and Gimeno (2003) show that the winter and early spring NAO type atmospheric circulation influences the extent of snow cover and the snow cover affects the atmosphere from late spring to early autumn leading to a mechanism that seems to be responsible for the multi-annual NAO persistence in the last half century. Gong et al. (2007) find that the Eurasian snow cover in autumn can modulate the winter Arctic Oscillation/North Atlantic Oscillation via stationary wave-mean flow interaction throughout the troposphere and stratosphere.

Several studies suggest that both the oceanic wind forced gyre circulation and the thermohaline circulation can actively interact with atmospheric flow to produce coupled decadal and interdecadal climate variability. In the paper of Grötzner et al. (1998) a decadal climate cycle in the North Atlantic that was derived from an integration with a coupled ocean-atmosphere general circulation model is described. The decadal mode shares many features with the observed decadal variability in the North Atlantic. The decadal mode is based on unstable air-sea interactions and must be therefore regarded as an inherently coupled mode. It involves the subtropical gyre and the North Atlantic Oscillation. The memory of the coupled system, however, resides in the ocean and is related to horizontal advection and to the oceanic adjustment to low-frequency wind stress curl variations. Although differing in details, the North Atlantic decadal mode and the North Pacific mode described by Latif and Barnett (1996) are based on the same fundamental mechanism: a feedback loop between the wind driven subtropical gyre and the extratropical atmospheric circulation.

Other studies have suggested a coupled mode of variability involving the thermohaline circulation. Modelling results of Timmermann et al. (1998) suggested that an anomalous strong thermohaline circulation produces positive SST anomalies over the North Atlantic. The atmospheric response is a strengthened NAO. The stronger NAO produces anomalous fresh water fluxes, and Ekman transport off Newfoundland and the Greenland Sea. This results in a reduced sea surface salinity which is advected by the subpolar gyre. This finally reduces the convective activity south of Greenland, thereby weakening the strength of the thermohaline circulation. This results in a reduced poleward oceanic heat transport and the formation of negative SST anomalies, which completes the phase reversal and results in multi-decadal variability.

### 5.5. NAO projections

Many future simulations project some decrease in the Arctic surface pressure in the 21st century, as seen in the

multi-model average. This contributes to an increase in indices of the NAO. In the recent multi-model analyses, more than half of the models exhibit a positive trend in the NAO (Rauthe et al., 2004; Osborn, 2004; Kuzmina et al., 2005; Miller et al., 2006). Although the magnitude of the trends shows a large variation among different models, Miller et al. (2006) find that none of the 14 models exhibits a trend towards higher Arctic SLP. In another multi-model analysis, Stephenson et al. (2006) show that of the 15 models able to simulate the NAO pressure dipole, 13 predict a positive increase in the NAO index with increasing CO<sub>2</sub> concentrations, although the magnitude of the response is generally small and model dependent. However, the multi-model average from the larger number (21) of models indicates that it is likely that the Northern Annular Mode (NAM) index<sup>13</sup> would not notably decrease in a future warmer climate. The average of IPCC-AR4 simulations from 13 models suggests the increase of the NAM index becomes statistically significant early in the 21st century.

The spatial patterns of the simulated SLP trends vary among different models, in spite of close correlations of the models leading patterns of interannual (or internal) variability with the observations (Osborn, 2004; Miller et al., 2006). However, at the hemispheric scale of SLP change, the reduction in the Arctic is seen in the multi-model mean, although the change is smaller than the inter-model standard deviation. Besides the decrease in the Arctic region, increases over the North Pacific and the Mediterranean Sea exceed the inter-model standard deviation; the latter suggests an association with a north-eastward shift of the NAOs centre of action (Hu and Wu, 2004). The diversity of the patterns seems to reflect different responses in the Aleutian Low and depends on whether or not additional sulphate aerosol forcing is imposed (Rauthe et al., 2004).

Yamaguchi and Noda (2006) discuss the modelled response of ENSO versus Arctic Oscillation (AO), and find that many models project a positive AO-like change. In the North Pacific at high latitudes, however, the SLP anomalies are incompatible between the El Niño-like change and the positive AO-like change, because models that project an El Niño-like change over the Pacific simulate a non-AO-like pattern in the polar region. As a result, the present models cannot fully determine the relative importance of the mechanisms inducing the positive AO-like change and those inducing the ENSO-like change, leading to scatter in global warming patterns at regional scales over the North Pacific. Rauthe et al. (2004) suggest that the effects of sulphate aerosols contribute to a deepening of the Aleutian Low resulting in a slower or smaller increase in the AO index.

Analyses of results from various models indicate that the NAM can respond to increasing greenhouse gas concentrations through tropospheric processes (Fyfe et al., 1999; Gillett et al., 2003; Miller et al., 2006). Greenhouse gases can also drive a positive NAM trend through changes in the stratospheric circulation, similar to the mechanism by which volcanic aerosols in the stratosphere force positive annular changes (Shindell et al., 2001). Models with their upper boundaries extending farther into the stratosphere exhibit,

<sup>13</sup> Please note that we use NAO, AO and NAM synonymously. For more information about the NAM see Thompson and Wallace (2001).



on average, a relatively larger increase in the NAM index and respond consistently to the observed volcanic forcing (Miller et al., 2006), implying the importance of the connection between the troposphere and the stratosphere. A plausible explanation for the cause of the upward NAM trend simulated by the models is an intensification of the polar vortex resulting from both tropospheric warming and stratospheric cooling mainly due to the increase in greenhouse gases (Shindell et al., 2001; Sigmond et al., 2004; Rind et al., 2005a). The response may not be linear with the magnitude of radiative forcing (Gillett et al., 2002) since the polar vortex response is attributable to an equatorward refraction of planetary waves (Eichelberger and Holton, 2002) rather than radiative forcing itself. Since the long-term variation in the NAO is closely related to SST variations (Rodwell et al., 1999), it is considered essential that the projection of the changes in the tropical SST (Hoerling et al., 2004; Hurrell et al., 2004) and/or meridional gradient of the SST change (Rind et al., 2005b) is reliable.

In summary, the future changes in the extratropical circulation variability are likely to be characterised by increases in positive phases of the NAM. The response in the NAM to anthropogenic forcing might not be distinct from the larger multi-decadal internal variability in the first half of the 21st century. The positive trends in annular modes would influence the regional changes in temperature, precipitation and other fields, similar to those that accompany the NAM in the present climate, but would be superimposed on the global-scale changes in a future warmer climate.

## 6. The impact of arctic sea ice changes on the NAO: observations and idealised experiments

### 6.1. Observed arctic sea ice changes and their impact on the NAO

The sea-ice cover impacts the heat-fluxes from the ocean to the atmosphere (see section “Arctic Sea Ice”) and therefore is anticipated to influence atmospheric circulation patterns and weather patterns (Francis et al., 2009). Factors driving the observed sea-ice variability have been investigated (e.g. Deser et al., 2000; Vinje, 2001; Deser et al., 2007), but the observed impact of sea-ice changes on the atmospheric circulation has received relatively little attention (see Francis et al., 2009). One reason is that to zero order the atmosphere forces the sea-ice anomalies (Deser et al., 2000; Wu and Zhang, 2010). Another reason might be that the observed atmospheric circulation anomalies are the result of many processes/forcing (Alexander et al., 2004; Budikova, 2009), and therefore it is more problematic to identify the impact of sea-ice anomalies on the atmosphere in observations.

A relationship between low Arctic sea-ice minima in autumn and late wintertime NAO was identified by Honda et al. (2009) in an observational analysis. Significant zonally elongated cold anomalies from Europe to the Far East in late winter are related to the decrease of the Arctic sea-ice cover in the preceding summer–autumn seasons. The cold temperatures were associated with negative NAO-like anomalies in the sea level pressure field. The delayed autumn/winter negative NAO-like pattern as a response of a reduced summer/autumn Arctic sea-ice cover is confirmed by Francis

et al. (2009); Wu and Zhang (2010). The relationship between a reduced Arctic sea-ice cover and the negative phase of the NAO is also established by modelling studies which is addressed in more detail in the following section.

### 6.2. Arctic sea ice changes in idealised experiments and their impact on the NAO

As pointed out in the previous section, numerous forcings are acting on the observed atmospheric circulation and it can be difficult to deduce changes in the observed atmosphere to sea-ice anomalies. The use of numerical models based on the equations of fluid motion and thermodynamics help to determine/confirm possible impacts of sea-ice changes on the atmosphere. Using model experiments where just the sea-ice boundary conditions are changed and every other external forcing is left unchanged makes it possible to relate the response to changes in the sea-ice.

One of the first sea-ice-sensitivity experiments was conducted by Newson (1973). In one integration the region of winter Arctic sea ice, defined as the mean climatological position, was replaced by open ocean maintained at freezing temperature. By comparing the two runs he found a marked warming in the lower Arctic troposphere. Not so obvious was a reduction in continental temperatures in the mid-latitudes. He found a southward displacement and reduction of the prevailing mid-latitude westerlies when there was no polar sea-ice cap. He concluded that the general decrease of the equator-to-pole temperature gradient reduces the strength of the westerly flow in mid-latitudes and that the atmospheric circulation becomes more blocked. Consequently, regions under the influence of the westerlies might become cooler. Similar results were obtained by Fletcher et al. (1973); Warshaw and Rapp (1973); and Royer et al. (1990).

Magnusdottir et al. (2004) carried out a series of experiments to understand how the variation in SST and sea-ice could affect the winter circulation using the atmospheric Climate Community Model 3 (CCM3). The study was constrained to the Atlantic sector and they used amplified anomalies in sea-ice and SST as forcing. They found out that anomalies in sea-ice extent are more efficient than SST anomalies at exciting an atmospheric response comparable to the amplitude to the observed NAO trend. They find that idealised sea-ice reductions in the Greenland Sea induce a negative response in the NAO in the winter season (see also Deser et al., 2004). The role of sea-ice anomalies in the Labrador Sea for forcing the NAO in January to March was emphasised by Kvamstø et al. (2004). Heavy sea-ice conditions in the Labrador Sea result in a significant negative NAO-like response, whereas reduced sea-ice extent causes a positive NAO response. Alexander et al. (2004) reach similar conclusions from their modelling study using observed sea-ice anomalies during December to February. They find in particular that the response to sea-ice reductions east of Greenland projects on the negative NAO phase, while enhanced sea-ice extend west of Greenland resembles the negative NAO phase.

Deser et al. (2007) analysed the development of the atmospheric response to Arctic sea-ice anomalies using the NCAR Community Climate Model 3.0 and the same sea-ice forcing as in Magnusdottir et al. (2004); Deser et al. (2004),

with the exception that they have performed 240 pairs of experiments to compute daily ensemble means. They find a baroclinic response in the vicinity of the forcing during the first stage of the response. The baroclinic structure reaches a maximum amplitude in approximately 5–10 days and persists for 2–3 weeks. Following the initial baroclinic stage of adjustment, the response becomes progressively more barotropic and increases in both spatial extent and magnitude. The equilibrium stage of adjustment is reached in 2–2.5 months and is characterised by an equivalent barotropic structure that resembles the NAO pattern. The maximum amplitude of this response is 2–3 times larger than that of the initial response.

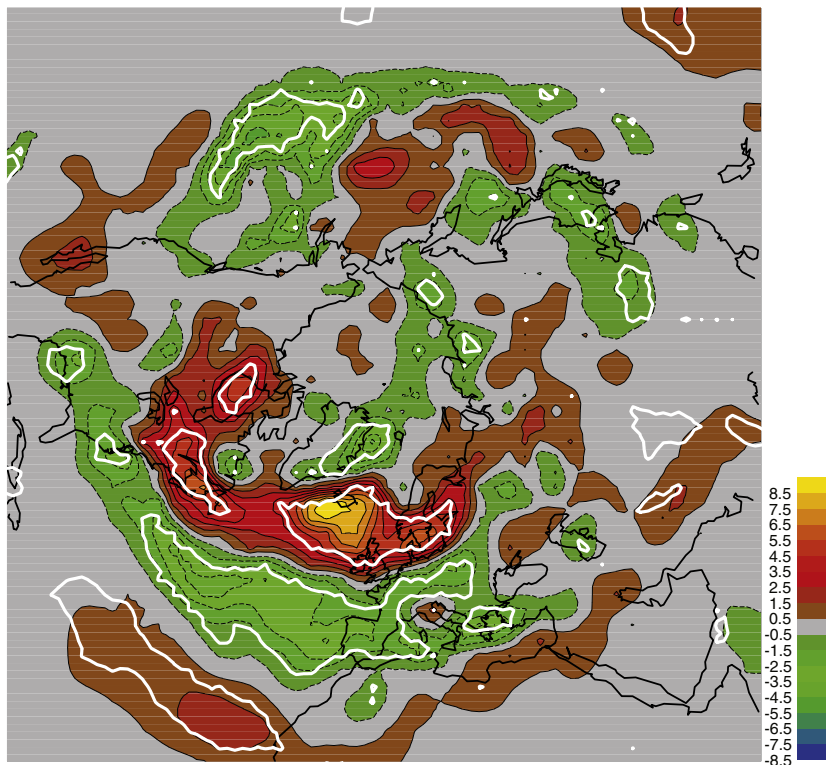
Seierstad and Bader (2009) find a strong seasonality in the NAO-response when they investigate the impact of a projected future reduction in the Arctic sea-ice cover on the NAO. Only in late winter (March) a clear NAO-like response was triggered suggesting that the background state can result in different atmospheric responses. The seasonality of the response is confirmed by Deser et al. (2010). Deser et al. (2010), using a similar set up as Seierstad and Bader, (2009), find a significant large-scale atmospheric circulation response during winter, with a baroclinic vertical structure over the Arctic in November to December and an equivalent barotropic in January to March. The equivalent barotropic response resembles the negative phase of the North Atlantic Oscillation in February only.

In contrast to Seierstad and Bader (2009); Deser et al. (2010), Singarayer et al. (2006) did not find a negative NAO response, when they forced the Hadley Centre Atmospheric model with observed 1980–2000 sea-ice extent and with projected future sea-ice reductions through 2100.

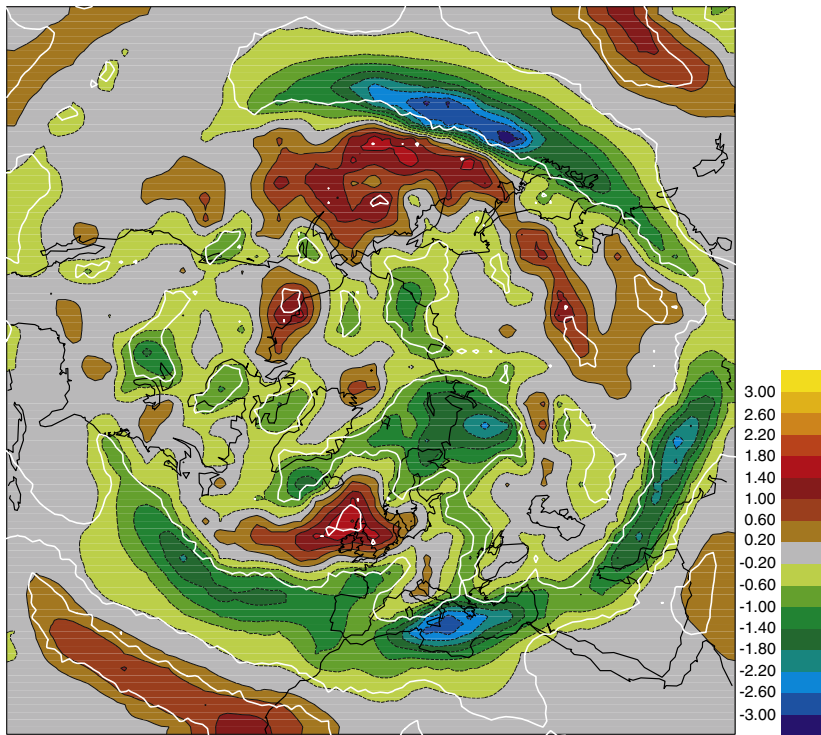
## 7. Storms and the NAO

The difference in storm-track density for the NAO for the NH winter computed from the 850 hPa relative vorticity from the NCEP–CFRS re-analysis is shown in Fig. 7. Positive and negative phase statistics are computed using the weighting scheme described in the appendix of Bengtsson et al. (2006). Fig. 7 shows a clear dipole-like structure for the track density in the Atlantic sector. The positive phase of the NAO is associated with an enhanced storm-track density in the northern North Atlantic and a reduced track density to the south.

The interaction between teleconnection patterns and storm tracks are twofold: the slowly varying component of the atmospheric circulation may influence storm track characteristics such as their intensity (e.g.: through the dipolar western Pacific and western Atlantic patterns) (Lau, 1988; Barry and Carleton, 2001), their count (Mailier et al., 2006), their “storminess” activity (Seierstad et al., 2007) and their north–south displacement (e.g.: associated with the Pacific/North American, eastern Atlantic and North Atlantic teleconnection patterns) (Lau, 1988; Hurrell et al., 2003).



**Fig. 7.** Difference in track density for the NAO for the NH winter computed from the 850 hPa relative vorticity from the NCEP–CFRS re-analysis for 1979–2009 and based on the CPC NAO index. Positive and negative phase statistics are computed using the weighting scheme described in the appendix of Bengtsson et al. (2006). Densities are number density per month per unit area, where the unit area is equivalent to a 5 degree spherical cap ( $\approx 1 \times 10^6 \text{ km}^2$ ). The white lines indicate regions where the p-values are below 0.05, i.e. significant at 95% computed using a Monte-Carlo method (Hodges, 2008).



**Fig. 8.** Projected change in the track density for the NH winter based on the 850 hPa relative vorticity from the ECHAM5 T63 coupled simulations for AR4 for the A1B scenario of anthropogenic forcing. Densities are number density per month per unit area, where the unit area is equivalent to a 5 degree spherical cap ( $\approx 1 \times 10^6 \text{ km}^2$ ). The white lines indicate regions where the p-values are below 0.05, i.e. significant at 95% computed using a Monte-Carlo method (Hodges, 2008). Figure taken from Bengtsson et al. (2006).

Rogers (1997) suggests that the NAO may be closely linked to the latitudinal aspects of storm track variability in the central Atlantic, but the low-frequency teleconnections that are linked to the predominant mode of the storm track variability in the North Atlantic are in the far northeastern Atlantic.

On the other hand, storm tracks may be responsible for triggering changes in teleconnection patterns such as the NAO. Studies have shown that even single storm events may act to shift the phase of the NAO (Rivière and Orlanski, 2007). Such a shift in the NAO phase may be achieved through Rossby wave breaking linked to the storm track activity (Vallis et al., 2004; Benedict et al., 2004; Strong and Magnusdottir, 2008): The NAO phenomenon is characterised by a meridional displacement of the upper tropospheric eddy-driven jet where positive and negative phases correspond respectively to a jet located further to the north and further to the south than usual (Rivière and Orlanski, 2007; Wettstein and Wallace, 2010; Athanasiadis et al., 2010).<sup>14</sup> Essentially two physical processes cause jet streams:

- conservation of angular momentum
- eddy momentum forcing associated with transient eddies

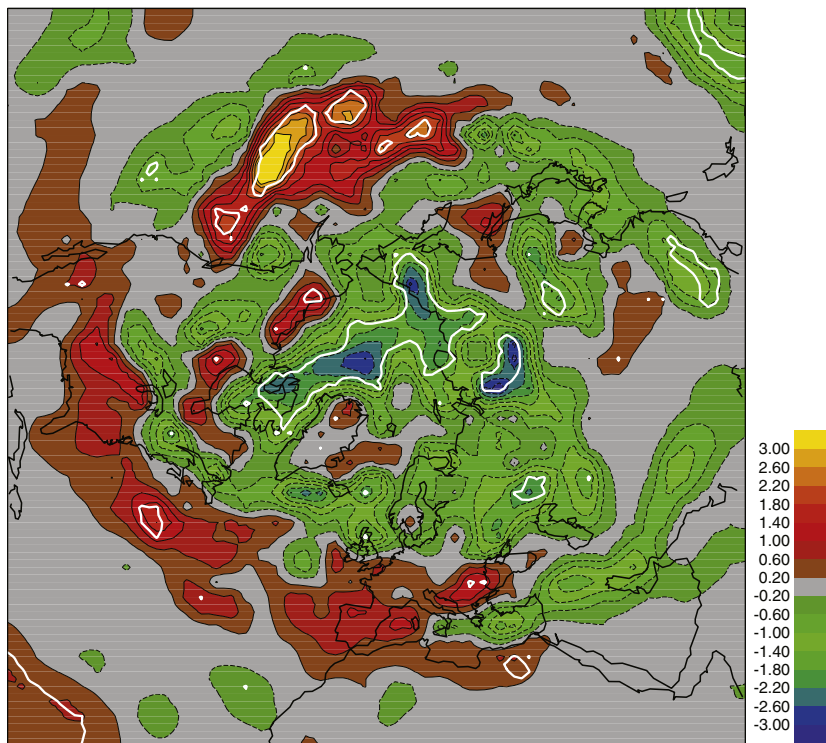
Because of these two mechanisms, it is common to refer to two different jet streams, named the subtropical and eddy-driven jets respectively. In many instances there is no spatial separation between the two jets, but even then the conceptual separation is still useful in many regards. The sub-tropical jet arises due to the westerly acceleration associated with poleward moving air in the upper branch of the tropical Hadley circulation. As the air moves north it comes closer to the rotation-axis. To conserve its angular momentum the parcel gets an eastward acceleration. This results in a westerly jet stream at upper levels at the poleward edge of the Hadley cell (Woollings et al., 2010): the subtropical jet.

The second mechanism is the momentum forcing arising from the effects of transient mid-latitude eddies like low and high pressure systems. By neglecting dissipation and vertical advection, and by approximating the Coriolis parameter by a constant value  $f_0$ , the quasi-stationary part of the zonal momentum equation for a non divergent flow can be written as (for details see Rivière and Orlanski, 2007):

$$\frac{\partial \bar{u}}{\partial t} + \bar{u} \frac{\partial \bar{u}}{\partial x} + \bar{v} \frac{\partial \bar{u}}{\partial y} = f_0 \bar{v}_a - \frac{\partial (\overline{u'^2})}{\partial x} - \frac{\partial (\overline{u'v'})}{\partial y}, \quad (1)$$

where overbars and primes denote respectively the quasi-stationary and the transient part. The last two terms on the right hand side show the convergence of the eddy momentum fluxes. Convergence of eddy momentum accelerates the mean flow, whereas divergence decelerates the flow. The jet

<sup>14</sup> We would like to point out that more recent studies show that the NAO and the East Atlantic (EA) pattern together resolve better changes in the latitude and speed of the jet (see e.g. Seierstad, 2008; Woollings et al., 2010).



**Fig. 9.** Track density difference for the NH winter computed from the 850 hPa relative vorticity from simulations of the ECHAM5 AGCM at spectral resolution of T213 for the current climate without sea ice and with observed sea ice. Densities are number density per month per unit area, where the unit area is equivalent to a 5 degree spherical cap ( $\approx 1 \times 10^6 \text{ km}^2$ ). The white lines indicate regions where the p-values are below 0.05, i.e. significant at 95% computed using a Monte-Carlo method (Hodges, 2008).

displacement characteristic of the NAO phenomenon depends strongly on the dynamics of the synoptic-scale waves and the way they break. Positive and negative phases of the NAO are closely related to anticyclonic and cyclonic Rossby wave breaking, respectively. When anticyclonic wave-breaking occurs the waves are tilted along the southwest–northeast direction which leads to a positive eddy momentum flux  $\overline{u'v'} > 0$  (Holton, 2004; Rivière and Orlanski, 2007). North and south of the wave-breaking the eddy momentum fluxes decrease. Hence, the eddy-momentum divergence  $\frac{\partial(\overline{u'v'})}{\partial y}$  is positive south of the wave-breaking maximum and negative north of the wave breaking maximum. As a consequence, the quasi-stationary zonal wind  $\bar{u}$  will increase in the north and decrease in the south. Anticyclonic wave-breaking will move the jet further north (see Rivière and Orlanski, 2007). For the cyclonic wave-breaking case similar arguments show that the jet moves southward.

The fact that high-frequency eddies with periods less than 10 days are important for driving the NAO growth has also been shown by Feldstein (2003). This is possible since the duration of an NAO anomaly is typically two weeks (Feldstein, 2003) and teleconnections can be produced by stochastic stirring that mimics baroclinic eddy development (Vallis et al., 2004).

Donat et al. (2010) demonstrate that storm events in Central Europe occur primarily during a moderate positive phase of the NAO, while strong positive NAO phases, occurring 6.4% of all

days, account for more than 20% of the storms. On a statistical average, during about 10% of the days with a strong positive NAO, a storm emerges over Central Europe.

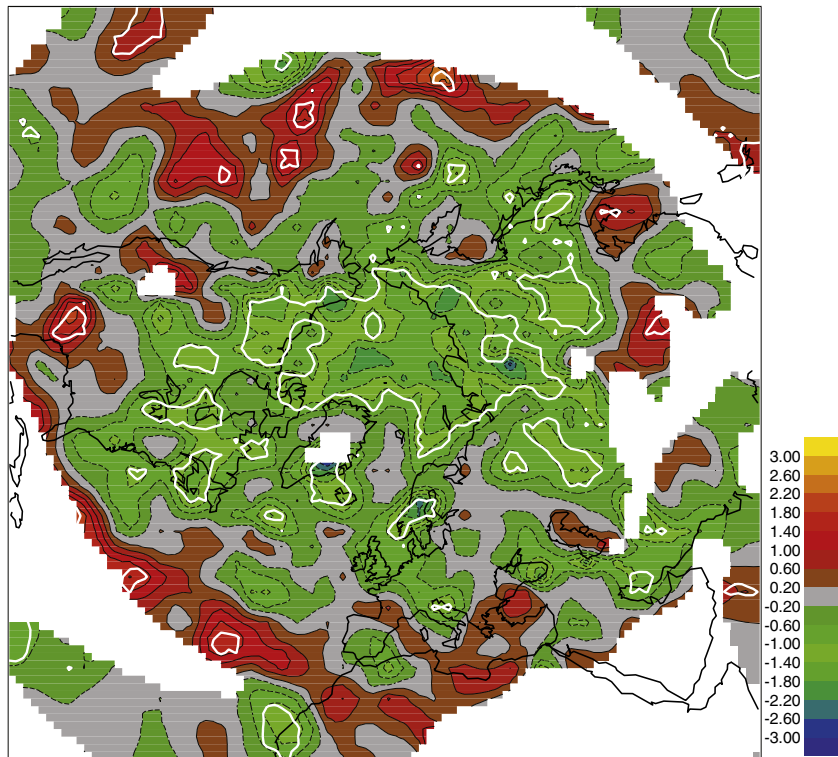
## 8. Conclusion

We have reviewed published literature concerning the impact of sea ice variability on the atmospheric circulation with a focus on storm tracks and the NAO variability. Sea ice has decreased substantially in recent years and it is expected to continue its decline. Observational data have shown:

- a significant reduction in Arctic sea-ice extent in all seasons ( $\approx 4\%$  per decade) (Serreze et al., 2007; Serreze and Stroeve, 2008)
- the largest negative trends ( $\approx 9\%$  per decade) at the end of the summer (September)
- a reduction of perennial sea ice
- a recent record low September sea-ice extent minima in 2007, and near records in 2008–2010
- the observed decrease in sea-ice extent is faster than the model predictions (Stroeve et al., 2007).

The simulated future projections can be summarised as:

- the sea-ice decline will continue over the 21st century and half of the model population exhibit ice-free summer Arctic by 2100 (Arzel et al., 2006).
- some studies project summer ice-free Arctic by 2040 (Holland et al., 2006; Wang and Overland, 2009)



**Fig. 10.** Mean cyclone speed difference for the same simulations as used for Fig. 9. The white lines indicate regions where the p-values are below 0.05, i.e. significant at 95% computed using a Monte-Carlo method (Hodges, 2008).

Will the sea-ice reduction have an impact on storms in the future?

Observations for the Arctic (Simmonds and Keay, 2009) show that the sea-ice variability has an impact on the strength of the storms: less sea ice leads to stronger storms. Simulations show that an overall reduction in the number of Arctic winter storms and a northward shift of mid-latitude winter storms in the Pacific are associated with a reduced Arctic sea-ice cover.

The future projected changes in storms may be summarised as:

- overall reduction in the number of arctic winter storms (see also Fig. 8)
- poleward shift of mid-latitude storm tracks in IPCC-type experiments.
- strengthening of the storm track north of the British Isles
- uncertainty over changes in the intensity of storms which depends on the definition of intensity. Precipitation will increase, but recent studies indicate no apparent change in extreme winds over the NH. Larger changes occur regionally.

The poleward shift of mid-latitude storms, both for the observations and future projections, is the most agreed on result. When it comes to the number or intensity of mid-latitude storms, results are still unclear. The disagreement among the studies might be related to: type of tracking algorithm, variable used for tracking, model resolution and the region selected for averaging storm statistics. It might also be due to differences in projected polar surface and upper

tropical troposphere warming, and changes in the MOC. These can produce different storm track responses.

For the average over the hemisphere, an increase in the number of extreme cyclones is found only when “extreme” is defined in terms of core pressure, while there is no change or a decrease in several models when defining “extreme” from the Laplacian of surface pressure, winds or vorticity around the core.

Will the Arctic sea-ice loss be a major driver for the NAO in the future?

Several observational and idealised sea-ice-sensitivity simulations indicate a delayed negative NAO-like response in autumn/winter to a reduced Arctic sea-ice cover (at least in some months). Most coupled ocean–atmosphere models of the last IPCC AR4 forced by the SRESA1B scenario predict a significant future reduction in Arctic sea-ice and a moderate tendency to a positive phase on the NAO at the end of the 21st century. This shows that the forcing of the sea-ice reduction is not the dominant forcing on the NAO in the future. There might be other processes that counteract the impact of the sea-ice loss on the NAO. We can only speculate, but one candidate could be the tropical ocean warming which has been shown to induce a positive NAO phase (Hoerling et al., 2001; Bader and Latif, 2003; Schneider et al., 2003; Hoerling et al., 2004; Hurrell et al., 2004; Bader and Latif, 2005). The Arctic sea-ice loss might dampen the increase of the NAO-index induced by other mechanism and provide a negative feedback mechanism in the Atlantic sector in the future. A negative feedback of the sea-ice on the atmosphere in the Atlantic basin is indicated by Alexander et al. (2004). Further studies are needed to confirm a negative

Arctic sea-ice feedback on the atmosphere in the Atlantic region and its importance for the future.

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