Recent Advances in understanding the Arctic Climate System State and Change from a Sea Ice Perspective: a Review

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Abstract

The sea ice is the central component and sensitive indicator of the Arctic climate system. The depletion and areal decline of the Arctic sea ice cover, observed since the 1970's, have accelerated after the millennium shift. While a relationship to global warming is evident and is underpinned statistically, the mechanisms connected to the sea ice reduction are to be explored in detail. Sea ice erodes both from the top and from the bottom. Atmosphere, sea ice and ocean processes interact in non-linear ways on various scales. Feedback mechanisms lead to an Arctic amplification of the global warming system. The amplification is both supported by the ice depletion and is at the same time accelerating the ice reduction. Knowledge of the mechanisms connected to the sea ice decline has grown during the 1990's and has deepened when the acceleration became clear in the early 2000's. Record minimum summer sea ice extents in 2002, 2005, 2007 and 2012 provided additional information on the mechanisms. This article reviews recent progress in understanding of the sea ice decline. Processes are revisited from an atmospheric, ocean and sea ice perspective. There is strong evidence for decisive atmospheric changes being the major driver of sea ice change. Feedbacks due to reduced ice concentration, surface albedo and thickness allow for additional local atmosphere and ocean influences and self-supporting feedbacks. Large scale ocean influences on the Arctic Ocean hydrology and circulation are highly evident. Northward heat fluxes in the ocean are clearly impacting the ice margins, especially in the Atlantic sector of the Arctic. Only little indication exists for a direct decisive influence of the warming ocean on the overall sea ice cover, due to an isolating layer of cold and fresh water underneath the sea ice.

1.Introduction

Sea ice is the central indicator of the state of climate in the central Arctic. Its sensitivity integrates changes in response to global scale climate forcing as well as of climate variability internal to the global climate system and internal to the Arctic. Sea ice is affected by thermal, radiative and dynamical changes of both Arctic atmosphere and ocean. Feedbacks from atmosphere and ocean are modifying the shape of the sea ice response.

Sea ice has been distinctly evolving since the start of satellite observations in 1979, which allow for an unprecedented accuracy in monitoring sea ice concentration and extent including interannual variability. A long term decline of summer sea ice extent of −12.9 % per decade is evident from the start of the record (Meier et al., 2012). After the year 2000, the decadal trend in summer sea ice extent loss has been strengthened and stands out as a period of distinct and persistent decline.

Before the start of the satellite era (1979), knowledge and observation of sea ice extent has long been either local or episodic. Reconstructions based on a limited number of local observations were carried out resulting e.g. in the HADISST2 data set (Rayner et al. 2006). Inconsistencies in the transition between traditional observations and the satellite record led to a recent correction of the sea ice extent time series before 1979 (Meier et al., 2012; Meier et al., 2013), showing large interannual variability on top of a rather stable summer sea ice extent during the 1950's – 1970's. The overall summer extent trend for the 1953–2011 period is estimated to −6.8 % per decade.

Modern knowledge on large scale sea ice thickness begins with submarine surveys, starting during the 1950's. Sonar measurements give a picture of thinning sea ice. Combining those with follow-up satellite retrievals from ICESat data (after 2003) gives an overall mean winter thickness decrease from 3.8 m in 1980 to 1.9 m in 2007-2008 (Kwok and Rothrock, 2009). The new generation Cryosat-2 satellite (Laxon et al. 2013) reconfirms the ice loss tendency.

For the time before 1950, knowledge on the state of Arctic climate is poor. The so called “early Arctic warming” observed during the 1930s and peaking during the 1940s can be clearly identified by atmospheric surface temperature anomalies from Arctic land stations (e.g. Johannessen et al, 2004). However, there is no known indication for an overall summer sea ice reduction. Reasons and mechanisms for the early Arctic warming are subject to discussion. It has been shown that natural variability likely contributed to the warming (Wood and Overland, 2010; Bengtsson et al., 2004). Hypotheses on dominating solar influences on the warm anomaly (e.g. Lean and Rind, 1998) could not be substantiated (Thejll and Lassen, 2000). Considering the millenium time scale, Kaufman et al. (2009), provide an extensive paleo-reconstruction of circumpolar land-based Arctic summer temperatures over the past 2000 years (based on proxies such as lake sediments, pollen records, diatoms, and tree rings), pointing out the recent Arctic warming as unprecedented during the last 2000 years.

As the globe is warming during recent decades, the Arctic is warming even stronger. A polar amplification of a global warming signal has first been envisaged by Arrhenius (1896) and later recognized by Broecker (1975). Manabe and Wetherald (1975) attributed the high-latitude amplification signal in one of the first coupled global climate models to what is known as ice-albedo-feedback. They also noted a role of the geographically different vertical structure of warming for the amplification, corresponding to the lapse rate feedback (see section 2.1). Recent research indicates a combination of various regional feedback mechanism in conjunction with circulation changes as reasons for observed and simulated Arctic amplification (Serreze and Barry, 2011; Pithan and Mauritsen, 2014). Arctic amplification is both reflecting and forcing sea ice changes.

The summer extent record after 2000 has turned into an amplified decline, eventually leading to a close series of summer record minima in 2002, 2005, 2007 and 2012. Those events are drastic illustrations of ongoing quantitative and qualitative changes. Especially the 2007 record sea ice minimum event marks a threshold in human consciousness of recent Arctic sea ice history (Nilsson and Döscher, 2013). The impact of Arctic processes became more obvious and a transformation of the Arctic climate system towards a “new Arctic” has been manifested, e.g. by the increased fraction of young first-year ice (Maslanik et al., 2011), thinner ice, warmer ocean and increased near-surface air temperatures. The “new Arctic” is expressing itself as a qualitative change noticeable not only by sea ice-related shifts, but also by enhanced meridional atmospheric circulation components (section 4.1) and warming of the Atlantic water layer in the mid-depth Arctic ocean, unprecedented in observed history (Spielhagen et al., 2011).

Detection of Arctic climate change in terms of atmospheric temperature has historically been difficult due to the regionally strong natural variability such as the early Arctic warming with a subsequent temporal cooling. Under such conditions, detection of a long term change signal or a trend requires long observation time series in order to prove significance. Only recently, a significant multi-decadal trend was possible to detect (Min et al., 2008), although human influence on sea-ice loss could actually have been detected as early as 1992 if currently used statistical methods (optimal detection analysis) had been available.

Our ability to attribute changes in various aspects of the Arctic climate increases when focusing on individual seasons. Anthropogenic signals have become detectable in colder seasons (Min et al., 2008). However, it is difficult to clearly attribute Arctic climate change to human influence based solely on observations (Overland and Wang, 2010). A strategy has therefore been to combine observation-based data and climate model data. In a recent study based on an up-to-date gridded data set of land surface temperatures and simulations from four coupled climate models, Gillet et al. (2008) concluded that anthropogenic influence on Arctic temperature is detectable and distinguishable from the influence of natural forcing, i.e. it is statistically attributable to human greenhouse gas emissions.

This conclusion and progress after previous studies was possible due to an updated gridded data set of land temperatures, allowing for more regional comparison with a model ensemble.

On this background of a detectable and attributable Arctic climate change, well visible in the sea ice cover, we find it useful to synthesize recent insights into the reasons for Arctic sea ice reduction and the underlying character of changes and the processes involved in the atmosphere and ocean. Recent reviews on the sea ice decrease (e.g. Stroeve et al., 2012 and Polyakov et al., 2012) take a specific look on a range of important contributing components. Here we attempt to add a wider system view on the sea ice decline, taking the changing overall Arctic physical climate system into account.

Arctic sea ice change includes global scale impacts, as well as regionally changing interaction mechanisms and trends. We give a review of existing peer-reviewed literature covering sea ice changes in combination with associated atmospheric and oceanic changes. Part of the reviewed work has been carried out during the international polar year (IPY) and the European DAMOCLES project. Special attention is given to recent updates of knowledge which sheds new light on previously existing results. We focus on the large-scale state and changes in the Arctic climate system affecting the sea ice cover and interacting with it, while recent advances in understanding small-scale physical processes were addressed in another DAMOCLES synthesis paper by Vihma et al. (2013). For discussion on consequences and impacts of a declining sea ice cover, see e.g. Meier et al. (2014). This introduction paves the ground by briefly summarizing the 20th century history of knowledge gain on the Arctic sea ice. Section 2 gives an overview of the Arctic climate system as an integral part of the global system. Section 3 gives a review of recent sea ice change, followed by section 4 on the influence of the atmospheric changes and section 5 on the impact of the ocean on sea ice change.

2. The Arctic as part of the coupled climate system

Climate change in the Arctic and on global scale are intensely intertwined. The Arctic represents a heat sink with both oceanic and atmospheric heat flux convergence. Our understanding is challenged by a range of interacting processes, complicated by a strong interannual and decadal variability in the Arctic climate. The recent Arctic warming in conjunction with sea ice depletion can be seen as part of and regional expression of a global warming. Arctic warming is detectable (Min et al., 2008) and can be statistically attributed to a globally changed atmospheric radiation balance due to increased atmospheric greenhouse gas concentrations (Gillet et al., 2008; Notz and Marotzke, 2012). The regional shaping and amplitude of the Arctic warming is governed by processes in the Arctic itself in conjunction with feedbacks which act differently within and outside the Arctic.

2.1 Arctic amplification

First climate model scenario simulations from the 1970's showed a global warming amplified in the Arctic (Manabe and Wetherald, 1975). Since then, an Arctic amplification of the global warming signal has been revealed in observations and turned out to intensify (Johannessen et al., 2004). Arctic amplification is now considered as an inherent characteristic of the global climate system (Serreze and Barry, 2011). A global scale warming triggers Arctic processes leading to a regionally amplified warming. The roles of retracting sea ice and snow coverage are widely described (e.g. Maksimovich and Vihma, 2012): The basic process of sea ice-albedo feedback starts to work in spring when the surface albedo decreases due to snow metamorphosis and melt. The feedback becomes even stronger when the melt exposes larger fractions of the ocean surface, and heat is effectively absorbed in the ocean (Perovich et al. 2007). This excess heat delays the start date of freezing with the consequence of thinner winter ice and a corresponding preconditioning of next summer's sea ice cover (Blanchard-Wrigglesworth et al., 2011). A corresponding process applies to the ice or snow surface under conditions of thinning and reducing multi-year ice. Decreasing sea ice albedo during the melting phase leads to thinner ice, memorized as long as into the following winter (Perovich and Polashenski, 2012; Notz, 2009). Those direct positive feedbacks in connection with reduction of ice concentration or thinning of ice explain that the strongest observed and projected future warming is located over the ocean/ice areas (Screen and Simmonds, 2010; Overland et al., 2011, Koenigk et al., 2011), with strongest seasonal signature in autumn and winter.

In addition to the role of sea ice-albedo-feedback, understanding of Arctic amplification became much more multifaceted during recent years, involving contributions of cloud and water vapour feedback, temperature feedback, atmospheric circulation feedbacks and reduced mixing in the Arctic atmospheric boundary layer, which all modifies the direct effects of Arctic climate warming (Soden et al., 2008). In addition, transport of heat into the Arctic by both ocean (e.g. Polyakov et al., 2010) and atmosphere (e.g. Serreze et al., 2009) plays a role.

The temperature feedback is commonly defined as the response to a warming of the surface or the atmosphere by increased longwave radiation by the 4^{th} power of the temperature. The effect is measurable at the top of the atmosphere. Due to generally colder temperatures in the Arctic, the increase of outgoing heat radiation in response to an equal temperature increase is less in Arctic latitudes, which potentially constitutes a contribution to the Arctic amplification.

The temperature feedback can be further refined and formally split into the Planck feedback, the contribution by a vertically homogeneous warming, and the lapse rate feedback. The latter, associated with the vertical structure of warming, builds on a reduced atmospheric lapse rate (“steepening”) under the conditions of a global warming (Soden et al. 2008), which leads to a greater warming in the upper troposphere than at the surface. The lapse rate in the vertical is affected by mixing, which in the tropics effectively conveys a surface warming signal to high altitudes, to be radiated to space. This is generally a negative feedback cooling the surface. In the Arctic however, the vertical transfer of heat is prevented by a stably stratified atmosphere, turning the lapse rate feedback regionally into a positive one, which contributes to the Arctic amplification.

Clouds and water vapour in the Arctic affect the regional radiation balance by blocking incoming short wave solar radiation, which gives a cooling effect on the surface. At the same time, increased downward long wave radiation is evoked with a warming effect on the surface temperature. In contrast to lower latitudes, Arctic clouds, especially low Arctic clouds, are found to warm the surface on an annual average (Kay and L'Ecuyer, 2013; Intrieri et al., 2002). The net effect of Arctic clouds thus constitutes an amplified warming in response to increased cloudiness, i.e. a positive cloud feedback. There is indication from various sources that Arctic cloud cover has increased during recent decades (see section 4.3)

The water vapour feedback refers to increased water vapour content in the atmosphere in response to a warming of the sea surface temperature (SST). Water vapour acts as a greenhouse gas and thus the water vapour feedback is generally positive, independent of location. Langen et al. (2012) broke down the impacts of the different feedbacks of Arctic amplification with the help of an idealized climate model configuration, with the result that the water vapour feedback does not in itself lead to an Arctic amplification. It does however strengthen the local response to other amplified positive feedbacks in the Arctic. Existing contributions to the Arctic amplification, such as by the ice-albedo-feedback and the combined temperature feedback, generate increased Arctic surface temperatures, which in turn increases water vapour emissions with an associated atmospheric warming in the Arctic.

The cloud feedback contribution is potentially capable of explaining an Arctic amplification on its own without the support of a sea ice albedo feedback. This is indicated in model studies with sea ice-albedo-feedback disabled by a fixed albedo (Langen and Alexeev, 2007; Graversen and Wang, 2009). Among the remaining mechanisms, the combined cloud feedback and the water vapour feedback (which not in itself generates an amplification) play the leading roles. Similar to the lapse rate feedback, the effect is supported by a generally stable stratification without convective mixing in the Arctic atmospheric boundary layer, hindering vertical mixing of humidity and thus keeping up increased humidity at lower levels. A more complete summary of the mechanisms involved in the Arctic amplification is given by Serreze and Barry (2011) and Pithan and Mauritsen (2014).

Important insights come from the analysis of Global Climate Model (GCM) ensembles, such as performed under the Climate Model Intercomparison Projects CMIP3 and CMIP5, and from individual climate models. Results do disagree on the ranking (the relative importance) of the different feedbacks. Given the finding of an Arctic amplification without any contribution by the sea ice-albedo-feedback (Langen and Alexeev, 2007; Graversen and Wang, 2009) we suggest that the different feedbacks might compete and take over when selected feedbacks are hampered in a self-adjusting process. According to the example above, the cloud feedback plays the leading role if the sea ice-albedo-feedback disabled. If the sea ice-albedo-feedback is active, it can dominate (Taylor et al., 2013).

Winton (2006) finds the Arctic amplification arising from “a balance of significant differences in all forcings and feed-backs between the Arctic and the globe”. Given that processes are implemented differently in various GCMs, diverse states of that balance are possible in principle, connected to different ranking of the feedbacks dominating the Arctic amplification, which might explain the spread in findings. Crook et al. (2011) and Taylor et al. (2013) suggest the surface albedo feedback as the largest contributor to the polar amplification. Taylor et al. (2013) emphasize that this is the case for the annual-mean and point at the cloud feedback beeing the second largest contributor to the Arctic amplification. Winton (2006) and Pithan and Mauritsen (2014) agree on a contributing but not dominating role of the surface albedo feedback. Pithan and Mauritsen (2014) find the largest contribution to Arctic amplification arising from the temperature feedback, followed by the surface albedo feedback as the second main contributor. Other contributions are found to be substantially smaller or even do oppose Arctic amplification.

While the regionally amplifying effects of sea ice-albedo-feedback, cloud, temperature and the water vapour feedbacks appear comprehensible, a current relevant question is to what extent those effects are triggered by regional processes only, or forced by changed transports of water vapour and heat via changed large scale circulation. There is indication that the regional Arctic amplification is enhanced by increased large scale heat transports into the Arctic as a dynamic response to the global scale water vapour feedback (Hansen et al. 2005). According to that hypothesis, water vapour transports are rearranged globally to even out the effect of the (positive) water vapour feedback in response to a warmer surface. The mechanisms involved are not understood, but a consequence of the hypothesized redistribution would be an inflow of water vapour into the Arctic. Model experiments (Langen et al., 2012, Boer and Yu, 2003) support this idea by analysing various feedbacks. Water vapour transports are found to change in a way that favours meridional patterns of response (Langen et al., 2012).

Evaluating the level of understanding of the Arctic amplification, we may conclude that reasonable concepts of the physics of the albedo, cloud, water vapour, temperature feedback and Planck feedbacks readily exist. Challenges remain, both in the quantification of the strength of the feedbacks and in understanding of the interactions between the various feedbacks. Clear indication exists though on the competition between diverse feedbacks that might lend varying importance to the various components under changing conditions. The Arctic amplification is maintained even if specific feedbacks are suppressed. This seems to ensure the existence of an Arctic amplification of atmospheric warming. For the sea ice this could mean a stable forcing towards less ice, even if the sea ice is a part of the competition among feedback processes. Realistic representation of the feedbacks in climate models is an ongoing complex task, as many of the feedbacks are related to subgrid-scale processes that need to be parameterized.

2.2 Coupled Arctic variability

Due to the Arctic's role as a heat sink with both oceanic and atmospheric heat flux components, changes of the large-scale northward heat transports must affect Arctic temperatures. Away from the surface, northward heat fluxes are less shaped by regional Arctic feedbacks. In the free troposphere away from the surface, Arctic temperature variations are highly determined by meridional heat flux anomalies. Yang et al. (2010) found a 50% (30%) contribution of positive (negative) atmospheric heat transport anomalies to decadal Arctic temperature trends based on reanalysis data in combination with microwave sounding estimates from polar-orbiting satellites covering the 1980s and 1990s.

Model results indicate that variability in northward heat transports into the Arctic in the ocean and atmosphere may compensate for each other. Ocean heat transport anomalies “modulate sea ice cover and surface heat fluxes mainly in the Barents Sea/Kara Sea region and the atmosphere responds with a modified pressure field” (Jungclaus et al., 2010), which results in an atmospheric transport anomaly of the opposite sign. The compensation mechanisms are not active at all times, and are connected to atmospheric circulation patterns in the Pacific sector of the Arctic, especially to the 2nd empirical orthogonal function (EOF) of the Pacific North-America Anomaly (PNA).

Anomalous atmospheric large-scale transports of atmospheric moisture have been found which support sea ice melt by enhancing long wave downward radiation. Effects of moisture transport are further described in sections 3 and 4.

The contribution of large scale ocean heat transport into the Arctic is discussed in section 5 of this review paper. In the Atlantic sector, a relation with the sea ice extent is well established (Koenigk et al., 2011; Holland et al. 2006), while direct impacts of Pacific inflow are difficult to prove.

Arctic sea ice variability and decadal scale changes can be generated both by regional Arctic processes (internally generated within the Arctic) or by global-scale forcing (externally forced by processes of global or hemispheric scale). Attempts to quantify the relative importance of both process types rely on climate model ensemble studies. Studies (Mikolajewicz et al., 2005; Döscher et al., 2010) suggest that the variability generated by the external forcing in recent climate is more important in most coastal regions than the internally generated variability. Both are, however, in the same order of magnitude and the relative importance varies locally within the Arctic. The degree of external vs. internal variability also depends on the state of large-scale atmospheric circulation. Northerly wind anomalies in the Atlantic sector of the Arctic support ice export and favour external control on the ice extent, likely due to external influence on the wind anomalies forcing the ice export.

Additional model studies point at strong internal variability during the summer (Dorn et al, 2012; Holland et al. 2011). Summer sea ice volume is significantly affected by the atmospheric circulation, which in turn is largely influenced by large scale atmospheric fields. Internal variability is particularly large in periods when the ice volume increases (Dorn et al, 2012).

3. Arctic sea ice state and change

3.1 Sea ice extent

Satellite-based observations of the Arctic sea ice extent exist since 1979. The 34-year record documents the seasonal and interannual evolution in the Arctic sea ice cover. Sea ice extent has decreased for all seasons, with strongest average decline for September of 84100 km^2 per year, and a moderate average decline during May of 33100 km^2 per year (Meier et al. 2013). After 1999 (1999-2010), the negative decadal trend of summer sea ice extent has been intensified to 154000 km^2 per year (Stroeve et al. 2012) and stands out as a period of persistent decline with record low September minima during 2002, 2005, 2007 and the latest record extent of 4.41 $10^6 km^2$ in September 2012. The latter four record events after 2000 are documented in Fig. 1, which shows the sea ice concentration together with the average ice margin for the years 1992-2006. The figure is provided by University of Hamburg and the SSM/I algorithms are described by Kaleschke et al. (2001).

Highest sea ice concentrations are found in the Arctic Ocean north of Greenland and in the Canadian archipelago as a result of prevailing winds across the Arctic. The summer ice extents from 2005 to 2012 were all lower than the minimum between 1979 and 2004. The ice reduction is characterized by a pronounced ice retreat within the East-Siberian, Chukchi and Beaufort Seas and in the Barents and Kara Sea. (Lindsay and Zhang, 2005; Comiso, 2006; Cuzzone and Vavrus, 2011). The shape of the remaining sea ice cover varies between the different record events. Since the late 1990's the Northeast passage is largely free of ice during September, with only small sea ice concentrations occurring e.g. in September 2007. Even the Northwest passage was largely ice free during September, starting 2007. Sea ice extent is also reducing during winter, mostly in the northern parts of the Barents Sea and in the northern North Pacific.

3.2 Sea ice thickness and volume

The accelerated decrease after 2000 is accompanied by changes in ice thickness, volume, albedo and sea ice age, which qualify for a regime shift towards a “new Arctic”, a term conceived after and inspired by the 2007 record sea ice low, referring to a qualitative change, with circumstances fundamentally different from 1980-2000 conditions (Comiso, 2006; Stroeve et al., 2007; Deser and Teng, 2008; Parkinson and Cavalieri, 2008; Liu et al., 2009).

Strong evidence exists for a decreasing Arctic sea ice volume, derived from occasional submarine-based upward-looking sonar observations. Thickness is measured in the central and western parts of the Arctic. The latest compilation by Rothrock et al. (2008) covers the period 1975 to 2000 and gives a winter mean ice thickness declining from a peak of 3.78 m in 1980 to a minimum of 2.53 m in 2000. This gives a decrease of 1.25 m until the year 2000. The mean annual cycle of sea ice thickness amounts to 1.12 m.

Altimeter equipped satellites, during the first years of this century (ICESat, 2003 – 2008), where capable to narrow the ice thickness with an uncertainty reaching locally 40-70 cm (Laxon et al., 2003; Kwok et al. 2009). Thin ice with less than 0.5 to 1 m in the marginal ice zone was excluded from analysis due to large uncertainties. Under those limitations, the winter sea ice thickness reduction from the submarine-based observations until year 2000 were extended to a thickness down to 1.89 m in 2008 (Kwok and Rothrock, 2009). Those values show an accelerated thickness loss after year 2000.

Estimates of overall Arctic sea ice volume have long been a challenge due to incomplete coverage of ice thickness data and its seasonal cycle. As a best guess approach, ocean-sea ice models, annually initialized with observed sea ice concentrations, can be used to infer sea ice volume. The Panarctic Ice Ocean Modeling and Assimilation System PIOMAS (Zhang and Rothrock, 2003) gives a trend over a 32 year period (1979-2011) of -2800 km^3/decade for October (Schweiger et al. 2011). Recent absolute volumes range between 28,700 km^3 in April and 12,300 km^3 in September. PIOMAS uncertainty is estimated to be 350 km^3 for October. Since the 1980's, the sea ice volume is reduced at a greater rate than the extent. By the mid-1990s, volume losses in September exceed ice extent losses by a factor of 4 in PIOMAS. Since then, volume/extent anomaly ratios approach smaller factors, arriving at a factor of about 2 at recently (Schweiger et al. 2011).

New satellite data from the European Space Agency CryoSat-2 (CS-2) mission allow for ice thickness estimates with a remaining uncertainty of 0.1 m in comparison with independent in-situ data when averaged over a large scale (Laxon et al. 2013). Starting 2011, sea ice volume loss over autumn and winter is about 500 km^3 per year (corresponding to 0.075 m per year in thickness), which fits well to peak thinning rates from the submarine-based observations. Between the ICESat (ending 2008) and CS-2 (starting 2011) periods, the autumn volume declined by 4291 km^3 and the winter volume by 1479 km^3 (Laxon et al. 2013). The seasonal cycle of volume loss and gain from CS-2 is greater than from PIOMAS. Longer term measurements by CS-2 will access better long term estimates of ice volume development.

Recent re-interpretation of ICESat data (2003-2008) obtains trends in sea ice volume of -1445 ± 531 $km^3\ a^{-1}$ in October/November and -875 ± 257 $km^3\ a^{-1}$ in February/March (Zygmuntowska et al. 2013). Taking into account algorithm uncertainties due to assumptions of ice density and snow conditions, the hypothesized decline in sea ice volume in the Arctic between the ICESat (2003–2008) and CryoSat-2 (2010–2012) periods may have been less dramatic (Zygmuntowska et al. 2013) than reported in Laxon et al. (2013).

The total annual sea ice volume budget is controlled by summer ice melt, wintertime ice accumulation, and the ice export. Naturally, those components of the volume budget depend on each other. As an example, ice growth increases material ice strength, which in turn reduces ice speeds. This potentially reduces the area of leads, which feeds back on ice growth.

Coupled atmosphere-ice-ocean numerical models are the principle tools to investigate sea ice volume budgets on the scale of seasons and years within the vast Arctic Ocean region. Derived from an ensemble of global climate models for recent climate conditions (1980-1999), a total melt of 1.1 m and an export of 0.2 m is balanced by 1.3 m of ice growth during the winter (Holland et al. 2010). Those figures largely agree with observation-based estimates derived from an Arctic heat budget combined with assumptions on latent heat of fusion and sea ice density (Serreze et al. 2007b).

Locally in the Beaufort Sea and around the North Pole, typical melting and growth rates have been about 20-50 cm per season each. That was the situation before the 2007 sea ice record minimum. During the 2007 event, Beaufort Sea bottom melting increased to about 200 cm (Perovich et al., 2008), which is explained by anomalously large fractions of open water, allowing for increased heat absorption by the ocean with subsequent lateral heat distribution underneath the ice.

For the climate since the year 2000, melt-export-growth imbalances grow. In the "Fourth Assessment Report of the Intergovernmental Panel on Climate Change" (AR4) global climate models largely agree on a decrease of ice volume resulting from increased annual melt during the melt season, rather than reduced growth during winter. This picture holds for the first half of the 21st century and is later reversed towards a dominance of reduced winter growth for the second half of the 21st century.

3.3 Sea ice age

Arctic sea ice is composed basically of the multi-year (perennial) and the first-year (seasonal) ice types. Sea ice thickness can be characterized by its age and the degree and type of deformation. The largest undeformed ice floe thickness is estimated to culminate at 1.5-2 m for the first-year ice and at

3-3.4 m for 7-9 year old ice-types. Pressure ridges can be as high as 20 m above the sea level, especially in coastal areas, but also in deeper areas such as the Beaufort Sea (Bourke and Garrett, 1987; Melling, 2002). Ridges can even grow larger under the water surface.

There is a good agreement on recent thinning between different data sources throughout the Arctic Ocean (Comiso et al., 2008; Kwok et al., 2009; Maslanik et al., 2011). This shrinking occurs primarily at the expense of the multi-year sea ice and thinning of ridged ice, while the thickness changes within the shifting seasonal ice zone are negligible (Rothrock and Zhang, 2005; Comiso, 2006; Nghiem et. al., 2007; Kwok et al., 2009). Among the multi-year ice types, the most extensive loss is seen for the oldest ice types. The fraction of total ice extent made up of multiyear sea ice in March decreased from about 75% in the mid 1980s to 45% in 2011, while the proportion of the oldest ice declined from 50% of the multi-year ice pack to 10%. By 2011, sea ice older than 5 years has almost vanished (Maslanik et al., 2011; from 2.8 10^3 km^2 in the 1980's to 0.4 10^3 km^2 in 2011). In terms of ice thickness, the mean value of the (former) perennial (now seasonal) ice zone was about 3-3.4 m during fall-winter season in 2003-2004, and approximately 2.3-2.8 m during 2007-2008 (Kwok et al., 2009). After summers with record low sea ice extent, the fraction of multi-year ice increases temporarily while the long term trend remains negative (Maslanik et al., 2011).

The major change in sea ice thickness distribution towards first-year ice is accompanied by a longer term decrease in the occurrence of thick pressure ridges in the central Arctic since the 1970's. Pressure ridges greater than 9 m (sum of ridge height and keel depth) showed a drop of 73%, as a result from comparing two older submarine missions in 1976 and 1996 (Wadhams and Davis, 2000). It is hypothesized that deep pressure ridges are more susceptible to bottom melting due to the large porosity of the deep ice material which allows for more efficient melting once the water warms (Amundrud et al. 2006, Wadhams, 2013). Despite local increase of ridge population due to increased ice moveability, there is a long term trend towards less deep ridges (Wadhams, 2013).

3.4 Sea ice motion

Arctic sea ice is constantly in motion under the effect of winds, ocean currents, tides, the Coriolis force, sea surface tilt and the internal resistance of the ice pack. The local air-ice momentum flux is usually the dominating forcing factor, and depends on the local wind speed, thermal stratification, and aerodynamic roughness of the surface. Under the stress sea ice floes crush, diverge and build-up pressure ridges. Recent changes in the ice drift have been mostly associated with changes in the internal resistance and atmospheric forcing; these effects are discussed below.

Arctic sea ice motion mirrors closely the background atmospheric circulation patterns (Inoue and Kikouchi, 2007). In winter a well developed Beaufort High in the western Arctic, and frequent and intense cyclonic motion in the eastern Arctic remove sea ice from the Siberian coast (Laptev, Kara and East-Siberian Seas) towards Greenland and the Fram Strait. In summer those transpolar winds and related ice drift speeds weaken. Day-to-day variability of surface winds are modulating the ice drift trajectories and velocities. Ice drift speeds range within 0-25 km per day (Zhao and Liu, 2007).

Interannual variability in the monthly mean ice drift has been attributed to the predominant atmospheric circulation patterns, such as the Arctic Oscillation (AO), the North Atlantic Oscillation (NAO), the Dipole Anomaly (DA; the second leading mode of sea-level pressure anomaly in the Arctic), and the Central Arctic Index (CAI). Wu et al. (2006) define the DA as a dipole anomaly corresponding to “the second-leading mode of EOF of monthly mean sea level pressure (SLP) north of 70°N … “. Earlier, Skeie (2000) found the second EOF of monthly winter SLP anomalies poleward of 30°N, named “Barents Sea anomaly”, to be highly influential on Eurasian climate. Overland and Wang (2010), referring to an analysis area north of 20°N, find a third EOF mode, which they called the Arctic Dipole (AD), reminiscent of the “Barents Sea anomaly” of Skeie (2000). Thus, the definitions of second or third modes vary. All versions commonly point at variability modes introducing meridional circulation components.

The close relationship of ice drift with the AO and NAO is well known (e.g., Inoue and Kikouchi, 2007; Kwok et al., 2009). Maslanik et al. (2007) suggested, however, that the AO is not a reliable indicator of the ice drift patterns that have favored sea ice decline in the western and central Arctic since the late 1980s. Also Zhang et al. (2008) suggested a decreasing control of the AO and NAO on the Arctic sea ice cover. The importance of the DA was demonstrated by Wu et al. (2006) and Wang et al. (2009). Recent work under the DAMOCLES project has, however, shown that over most of the Arctic the annual mean ice drift speed forcing is better explained by the CAI, calculated as the sea level pressure difference across the Arctic Ocean along meridians 270°E and 90°E (Vihma et al., 2012). The drift speed is more strongly related to the CAI than to the DA partly because the CAI is calculated across the Transpolar Drift Stream (TDS), whereas the pressure patterns affecting the DA sometimes move far from the TDS. CAI also has the benefit of being insensitive to the calculation method applied, whereas the DA, as the second mode of a principal component analysis, is sensitive both to the time period and area of calculations (Vihma et al., 2012). Arctic-wide, different combinations of atmospheric circulation indices (such as the CAI, DA and AO) explains 48% of the variance of the annual mean ice drift in the circumpolar Arctic, 38% in the eastern Arctic, and 25% in the Canadian Basin (Vihma et al. 2012).

Sea ice drift velocities have gradually increased since the 1950's. Significant positive trends are present in both winter and summer data (Häkkinen et al., 2008). The Arctic basin-wide averaged trend in drift speed between 1992 and 2009 has increased by 10.6% per decade (Spreen et al. 2011). The trend is strongest after 2004 with an average increase of 46% per decade. The drift of the sailing vessel Tara in 2006-2007 in DAMOCLES was almost three times faster than that of Fram in 1893-1896 (Fig. 2) along a similar path in the central Arctic (Gascard et al., 2008), but the contributions of various forcing factors to the difference is not quantitatively known. The winds at Tara were rather weak but their direction favoured the trans-polar drift (Vihma et al., 2008). The TDS has strengthened especially in summer between the late 1970s and 2007 (Kwok, 2009).

Considering the ice drift evolution from the 1950’s to 2007, Häkkinen et al. (2008) identify the primary reasons for the ice drift trend as increasing wind speed, related to increased storm activity over the TDS. Drift speed changes after the year 2000 are also connected to net strengthening of ocean currents in the Beaufort Gyre and the transpolar drift, propelled by a positive DA anomaly for the mean summer circulation (2001–2009), which also enhances summer sea ice export through the Fram Strait (Kwok et al. 2013).

Rampal et al. (2009) and Gimbert et al. (2012) find that the increase in drift speed since 1979 is rather related to a thinner sea ice with a reduced mechanical strength. Spreen et al. (2011) detected signs of both wind and ice thinning effects in 1992–2009 with the ice thinning likely more important. According to Vihma et al. (2012), atmospheric forcing cannot explain the increasing trend in drift speed in the period 1989–2009, but did explain a large part of the inter-annual variance, which cannot be explained by changes in ice thickness.

More information arises from recent reports on the impact of younger ice. Regionally, “positive trends in drift speed are found in regions with reduced multi-year sea ice coverage. Over 90% of the Arctic Ocean has positive trends in drift speed and negative trends in multiyear sea ice coverage” (Kwok et al., 2013). Changes in wind speed explain only “a fraction of the observed increase in drift speeds in the Central Arctic but not over the entire basin” (Spreen et al. 2011). In other regions, it is the ice thinning that is the more likely the cause of the increased ice drift speed.

Reviewing the above papers explaining increased ice drift speeds, points to an increasing importance of the effects of thinning and age for the more recent past, while increased wind speeds dominate before 1990.

A direct consequence of increased ice speeds is a temporally increased sea ice export through the Fram Strait (Kwok et al., 2013). Buoy data from 1979 to mid-1990s suggested an increasing trend in the ice area export via the Fram Strait, mostly due to a positive phase of the AO (Polyakov et al., 2012).

Increased ice movement is also contributing to specific events of rapid ice extent loss. During 2007, first year ice from the Chukchi Sea intruded the Northern Beaufort Sea. Combined with increased pole-ward summer ice transport from the western Arctic, a reduced fraction of multiyear ice provided ground for the 2007 record event (Hutchings and Rigor, 2012). Ice loss by Fram Strait export is stimulated by suitable local winds over the Fram Straits. Sea ice export variability is strongly determined by variations in the sea level pressure gradient across the Fram Strait. This finding is based on numerical simulations with a GCM (Koenigk et al., 2006), and supported by analysis of ice export observations in relation to atmospheric reanalysis (Tsukernik et al., 2010). Positive CAI and DA were observed during summer 2007, coinciding with an increased ice export (Zhang et al. 2008). Before 2007, between 1979 and 2006, no significant summer SLP forcing of Fram Strait ice motion was found. A generally increased Fram Strait ice area export on a decadal scale cannot be detected (Spreen et al., 2009). A slight increase in SLP pressure gradient, potentially forcing increased ice export, is compensated by a parallel decrease in the sea ice concentration (Kwok et al., 2009; Polyakov et al., 2012).

As the ice thins and is subject to increased weather impact, even the frequency of cyclones during late spring and summer is affecting the summer sea ice area. Low September sea ice areas are generally connected to below normal cyclone frequency during spring and summer over the central Arctic. Less cyclones means increased sea level pressure, enhanced anticyclonic winds, a stronger transpolar drift stream, and reduced cloud cover, all of which favour ice melt (Screen et al., 2011). Thus, storm activity over the central Arctic has a preconditioning effect on the outcome for the summer sea ice area and extent. An obvious question is whether the storm activity over that region has changed during the recent decades. Observations show a northward shift of storm tracks, which is discussed in further detail in section 4.

3.5 Snow and freezing/melting processes

Ice floes in winter are almost always covered by snow. The snow depth varies between 0-100 cm on horizontal distances of 10-100 metres, with no relationship to the ice type and ice thickness, except that in winter only thin, young ice in refrozen leads is free of snow (Walsh and Chapman, 1998; Perovich et al., 2002; Perovich and Richter-Menge, 2006; Gerland and Haas, 2011). Low thermal conductivity and high heat capacity of the snow explain the fact that the snowpack acts as a good insulator for the sea ice. In the presence of snow the response of the sea ice temperature to perturbations in air temperature is largely weakened.

Little is known about changes in snow thickness on top of sea ice. The most extensive snow information available is based on measurements made at the Russian drifting stations from 1954-1991 (Radionov et al., 1996) and airborne expeditions with landings on sea ice from 1937-1993, but there are no contemporary, systematic, basin-scale in-situ observations of snow thickness on top of Arctic sea ice. Snow thickness estimates based on remote sensing have been developed (Brucker et al., 2013), but they are not accurate over deformed ice and multi-year ice in general. On the basis of ERA-Interim reanalysis, Screen and Simmonds (2012) detected a pronounced decline in summer snowfall over the

Arctic Ocean between 1989 and 2009. It resulted from a change in the form of precipitation; snow turns into rain due to lower-tropospheric warming. This has resulted in a reduced surface albedo over the Arctic Ocean, which Screen and Simmonds (2012) estimated to be comparable in order of magnitude to the decrease in albedo due to the decline in sea ice cover. Thus, the decline in summer snowfall has likely contributed to the thinning of sea ice during recent decades.

Satellite retrievals of the spring onset of snow melt, from both passive and active microwave observations, demonstrate the long-term tendency towards earlier surface melt, with a mean of about 2.5 days per decade in the central Arctic (Markus et al., 2009), reaching locally 18 days per decade, in particular within the central western Arctic (Maksimovich and Vihma 2012). Concurrently, the fall freeze-up appears to be more and more delayed in the season (Markus et al., 2009), both within the open sea and on top of the sea ice that survived the melt season. These two essential processes, spring melt onset and fall freeze-up, affect the sea ice extent behaviour in time in a non-linear way, as well as thickness and the resulting volume (Maksimovich and Vihma, 2012). A few days of earlier surface melt initiation (typically occurring during May-June) drastically increase the absorption of solar energy, with the effect propagating through the entire melt season.

Radiation measurements in the central Arctic in combination with numerical experiments allowed to quantify the contribution of the earlier spring melt initiation and later fall freeze-up (Perovich et al., 2007). A one-day earlier spring melt corresponds to additional ice melt of 3 cm during the melt-season. In contrast, the fall freeze-up (typically occurring in late August – November) delayed by one day contributes to about 0.5 cm of summer ice melt in the same season. As a positive feedback, the earlier spring melt contributes to earlier ice thinning, further additional heat storage in the upper ocean during the melt season (Frey et al., 2011), and thus retarding the fall freeze-up (Armstrong et al., 2003; Gerdes, 2006; Perovich et al., 2007a,b). The spring melt initiation and the fall freeze-up timing are statistically related (Maksmovich, 2012), in particular in the Eastern Arctic Basin covered by first-year ice. The delayed ice formation plays a great role in the atmospheric warming during the early polar night season. As an example, the ocean heating of the lower atmosphere was nearly 3 times greater in September-November months during years with the exceptional ice retreat (2005-2007) compared to earlier years with larger summer ice extent (Kurtz et al., 2011).

The atmospheric thermodynamic forcing on sea ice thickness is transmitted via radiative and turbulent surface fluxes. Our knowledge of the climatology of radiative and turbulent fluxes is based on few observations only, the year-round SHEBA campaign being the most important one (Persson et al., 2002). The radiative fluxes are typically larger in magnitude than the turbulent fluxes. In winter, the upward longwave radiation exceeds the downward component; the negative longwave radiation on the snow surface is typically balanced by a downward sensible heat flux and heat conduction through the ice and snow. The latent heat flux is close to zero in winter. In summer, net shortwave radiation is the dominating flux, the net longwave radiation is less negative than in winter, latent heat flux is upwards, and the sensible heat flux may be either upwards or downwards (Persson et al., 2002). Unfortunately there are not enough observations available to estimate possible trends in the turbulent surface fluxes. For moisture fluxes, see section 4.3.

Albedo at the surface of sea ice or snow on top of sea ice is the crucial property limiting the effect of shortwave radiation on the ice (on the recent advances in physics and parameterizations, see Vihma et al. (2013)). Values for albedo at the ice or snow surface have long been derived from local direct observations. Improvements arise from satellite based algorithms, which even allow for accessing the long term temporal development of ice/snow albedo. Albedo trends during the 1980s and 1990s were rather weak compared to the trends after the mid 1990's (Wang and Key 2005). Laine (2004) finds a surface albedo trend for the Arctic Ocean close to zero, based on advanced very high resolution radiometer (AVHRR) Polar Pathfinder satellite observations for the years 1982 – 1998. Later, a long term decrease of the albedo in the sea ice zone has been detected (Riihelä et al. 2013) based on data products from the Satellite Application Facility on Climate Monitoring (CM SAF) covering 1982 - 2009. For the mean August sea ice zone (all surface areas with more than 15% sea ice concentration), a significant trend of -0.029 per decade has been found for the albedo (Riihelä et al. 2013). This includes even the effect of leads, which have a much lower albedo as any type of sea ice. Both increased lead areas and reduced ice surface albedo contribute to the trend.

Earlier timing of melt onset is an important influence on reduced sea ice albedo (see above). For comparison, simulated recent climate between 1982 and 2005 within the CMIP5 project gives a cross-model average albedo trend of -0.017 per 24 years (Koenigk et al, 2013), corresponding to -0.0071 per decade. This is about half of the observed trend. Climate models in CMIP5 however show large differences in albedo formulations and values.

Sea ice albedo depends on a range of influences (e.g. ice thickness, age, temperature, melt pond fraction, length of melting/freezing seasons and others). Melt ponds on the ice are reducing the sea ice albedo (Perovich et al. 2011). A quantification of Arctic-wide melt pond occurrence and effects requires satellite observations. Recent progress in algorithm development enables observations over complete melting periods. Anomalously high melt pond fractions are found during the summers of the record low sea ice years of 2007 and 2012, based on the Moderate Resolution Imaging Spectroradiometer (MODIS) satellite sensor (Rösel and Kaleschke, 2012). However, long term trends of melt pond fractions cannot be detected with statistical significance.

The important role of melt ponds on sea ice albedo is supported by numerical simulations of Arctic climate. Under recent climate conditions, melt ponds predominantly develop in the continental shelf regions and in the Canadian archipelago. Use of melt pond parameterizations, compared to classical albedo formulations without or only with simplistic recognition of melt ponds, leads to systematically reduced albedos, enhanced sea ice melt, reduced summer ice thickness and concentration (Karlsson and Svensson, 2013; Roeckner et al., 2012; Flocco et al., 2012) and contribute about 1 Wm^{-2} to forcing of ice melt (Holland et al., 2012).

Sea ice melt is further exacerbated by deposition of atmospheric aerosols (dust and soot) on the highly reflective snow and bare ice surface, reducing the surface albedo. In presence of soot, the absorption of solar radiation is more efficient and the internal heat storage is larger, supporting earlier and faster snow melt (Clarke and Noone, 1985; Grenfell et al., 2002). Black carbon is identified as the dominating absorbing impurity. The effect on climate forcing is estimated +0.3 W/m^2 in the Northern Hemisphere (Hansen and Nazarenko, 2004), to be compared with +0.6 W/m^2 overall global forcing by black carbon and a total of 2.3 W/m^2 (IPCC 5th assessment report) in anthropogenic radiative climate forcing.

GCM-based studies confirm the effect (Roeckner et al., 2012; Holland et al., 2012). Recently, the effects of soot on different ice types has been recognized. Given a background of black carbon on the ice, first year sea ice is more sensitive to black carbon additions compared to multi-year ice (Marks and King, 2013). The first year sea ice is scattering incoming radiation to a lesser degree than multi-year ice. This points to a positive feedback of the growing dominance of first-year ice, which facilitates stronger melting due to more efficient albedo reduction by black carbon. The knowledge situation is complicated by fresh snow covering the soot existing on the ice, thereby temporarily mitigating the effect of black carbon on sea ice.

We are witnessing an Arctic sea ice pack that is thinning, becoming younger and more moveable, with a decreasing albedo and lengthening melting season. All this makes the ice cover more susceptible to quick response to a warming climate. In that sense, the Arctic climate system has reached a new era with decreased stability of the ice cover.

3.6 Challenges in the understanding of sea ice evolution and sources of uncertainty

The understanding of the sea ice state variability and trends as described above, is challenged by the problem that information on changes in sea ice thickness is inaccurate, in particular for the summer period. Still much less is known about potential changes in snow thickness on top of sea ice. Key results, such as the findings by Screen and Simmonds (2012) on the decrease of snow fall and increase of rain over the Arctic Ocean, are based on reanalysis data, which cannot be completely verified by direct observations. A spatially and temporally extensive change from snowfall to rain may have more potential to reduce sea ice albedo than e.g. black carbon.

Further uncertainty arises from imperfect estimates of sea ice extent and concentration. Depending on the processing algorithm applied to the microwave satellite data, the Arctic sea ice extent may still have an uncertainty of up to 1 x 10^6 km^2 (Kattsov et al., 2010). The treatment of new, thin ice in refrozen leads is one of the factors generating scatter in the results. The generation of consistent time series over long periods is challenging because of the changes in the sensors onboard satellites (Cavalieri and Parkinson, 2012). Further, changes on ice type, level of fracturing, amount of superimposed ice, and areal coverage of melt ponds are not well known, but various new and anticipated satellite remote sensing products, and combined use of remote sensing and thermodynamic modelling, may soon provide improvement in the situation.

To assess an accurate mass and volume budget for Arctic sea ice, thickness information is essential. Published results on ice drift and export demonstrate a large inter-annual and decadal variability. The recent increase in ice drift speed is mostly due to ice becoming thinner and mechanically weaker. The effects of increased drift speed and decreased ice concentration have balanced each other so that there is no long-term trend in the ice area flux out of the Fram Strait. Hence, as also the ice thickness has decreased, the ice volume transport must have decreased. Despite of this, the relative importance of ice export in the mass balance of Arctic sea ice has not necessarily reduced, as the ice volume in the Arctic has decreased together with the volume transport.

Despite those uncertainties, the picture of the Arctic sea ice that becomes thinner and younger, and reduces in extent, is robust because the signal is strong and verified through different sources. However, understanding of specific mechanisms and budgets in detail is still vague. This is especially the case for the changing sea ice volume components and snow processes.

3.7 Future sea ice projection and prediction

Global climate models are tools supporting an integrated understanding of the Arctic climate system and its link the other geographical areas. Although imperfect by definition, models allow for process studies and future climate projections including assessment of uncertainty. Global climate models of the CMIP5 project, when run for observed periods, tend to underestimate the sea ice decline and differ greatly among each other (Massonet et al., 2012) (note: in contrast to climate prediction, those CMIP5 simulations are not initialized with recent observations and suffer from natural variability not necessarily in phase with reality). Identifying subsets among the simulations, those models with near-realistic atmospheric circulation can better simulate the sea ice extent decline after year 2000. However, many models suffer from a circulation bias. A large uncertainty is also seen in sea ice future projections. It is related to a generally too small decrease rate or too late sea ice drop. Reasons are to be seen in the different models' parametrizations and biases in atmosphere, ocean, ice and the coupling between those component models. Also model differences of sea ice albedo contribute to the large uncertainties in the Arctic climate as simulated by global climate models (Hodson et al., 2013), and results in large differences for the Arctic radiation balance (Karlsson and Svensson., 2013).

Future progress in the ability to simulate Arctic sea ice requires to better quantify heat exchange between sea ice and atmosphere/ocean and sea ice thickness. It will also be necessary to reduce model circulation biases.

Sea ice prediction (different from projection) on seasonal to decadal time scale requires careful initialization with ocean and sea ice conditions. Additional potential is seen in coupled initialization of land. When initialized climate models are run in ensemble mode (several runs differing slightly only in initial conditions), the spread of the results can be explored to assess the potential predictability of the Arctic, i.e. the upper limit of climate predictability on seasonal to decadal time scales. Sea ice thickness appears to be highly predictable along the ice edges in the North Atlantic Arctic Sector on decadal average (Koenigk et al., 2012), due to a strong correlation with the meridional overturning circulation in the North Atlantic Ocean. Such results give us a positive glimpse of possible future expectations to climate prediction in the Arctic.

4. The role of the atmosphere and its impact on sea ice

The atmosphere interacts with the Arctic sea ice decline via thermodynamic effects on ice melt and dynamic effects on ice drift (the latter discussed in Section 3.4). The direct thermodynamic atmosphere-sea ice coupling occurs via the radiative and turbulent surface fluxes. whereas precipitation has a strong indirect effect on this coupling via modification of radiative fluxes, surface albedo and snow thickness (Section 3.5). Meteorological observations over sea ice are limited, and direct measurements of surface fluxes and precipitation are extremely rare. Coastal observations are not representative for the sea ice zone. Radiative and turbulent surface fluxes from atmospheric reanalyses include large errors (Wesslen et al., 2013; Tastula et al., 2013) and the quality of reanalyses' precipitation over sea ice is poorly known (Jakobson and Vihma, 2010). Hence, much of our observationally-based knowledge on atmospheric-driven thermodynamic effects on sea ice decline originates from analyses of processes and variables that indirectly, rather than directly, affect sea ice melt and growth.

Among the relevant atmospheric conditions for Arctic sea ice change are the large-scale circulation patterns, characterized, among others, by the AO, NAO, and DA (as introduced in Section 3.4). Large-scale circulation patterns are inherently and interactively related to cyclone statistics and properties. Cyclones are responsible for a major part of the transport of heat and water vapour into the Arctic. Essential characteristics of the Arctic atmosphere also include cloud coverage and properties and the vertical structure of the atmosphere, from the atmospheric boundary layer (ABL) to the stratosphere.

4.1 Large-scale circulation and cyclones

Large scale oscillation patterns have been influential in preconditioning and forcing the observed sea ice decline at times. Both observational and modelling studies have demonstrated that the positive polarity of AO or NAO drove a decrease in sea ice extent or thickness between 1980 and the mid 1990s. Before 1980, the relation was less efficient because the NAO pattern was shifting in space around 1980 (Hilmer and Jung 2000). Such spatial shifts have been shown to impact on Arctic temperatures throughout the 20th century, characterized by varying angles of the axis between the NAO's centres of action (Jung et al., 2003; Wang et al., 2012). During the positive NAO/AO years after 1980, and especially during the most positive years 1989 – 1995, altered surface winds resulted in a more cyclonic ice motion and a more pronounced Transpolar Drift Stream (TDS) connected to enhanced ice openings, thinner coastal ice during spring and summer, and to increased sea ice export (Rigor et al. 2002; Serreze et al. 2007). The continued downward trend of sea ice extent after the mid 1990s are interpreted as delayed response in addition to other effects such as the ongoing increase of atmospheric temperatures (Lindsay and Zhang 2005). In the winter 2010/2011, a strongly negative AO was observed (Stroeve et al., 2011). Maslanik et al. (2011) argue that this explains a recent partial recovery of multiyear ice extent (see section 3.3).

During this century, the large-scale circulation in the Arctic has changed from a zonally dominated circulation type, which can be well characterized by the AO, to a more meridional pattern characterized by the AD, where a high-pressure center is typically located in the Canadian Arctic and a low in the Russian Arctic (Overland and Wang, 2010). This favours advection of warm, moist air masses from the Pacific sector to the central Arctic, contributing to the sea ice decline (Graversen et al., 2011) and rapid sea ice loss events (Döscher and Koenigk, 2013). Through increased release of ocean heat to the atmosphere during autumn, the sea ice decline has, in turn, contributed to a modification of large-scale atmospheric circulation, favoring a positive AD (Overland and Wang, 2010).

Another noteworthy aspect in recent large-scale circulation is that during the six latest years the strong Arctic warming has not been supported by positive values of the Pacific Decadal Oscillation (PDO) index (Walsh et al., 2011). The AO, DA/AD, and PDO closely interact with cyclone statistics. The cyclone activity is most vigorous in the Greenland Sea during all seasons, except summer, when the Norwegian, Barents and Kara Seas have a comparable amount of activity (Sorteberg and Walsh, 2008). The number of cyclones travelling into the Arctic is approximately similar in all seasons, but in winter the cyclones are more intense and shorter lived than during summer.

Approaches to Arctic cyclone statistics exit since the 1950 with very limited observations. More complete surveys were undertaken by e.g. Serreze (1993), and McCabe et al. (2001), revealing a positive trend of Arctic cyclone frequency for the period 1952 - 1997 for the winter.

More recent studies have addressed recent changes in synoptic-scale cyclones in the sub-Arctic and Arctic. A statistically significant increasing trend in the frequency of cyclones entering the Arctic during the recent decades has been detected e.g. by Zhang et al. (2004), Trigo (2006), Sorteberg and Walsh (2008), and Sepp and Jaagus (2011), suggesting a shift of cyclone tracks into the Arctic, particularly in summer. Analogously to synoptic-scale cyclones, Polar lows have migrated northward (Kolstad and Bracegirdle, 2008; Zahn and von Storch, 2010), which may be due to the retreating sea ice margin.

According to Sepp and Jaagus (2011), however, the frequency of cyclones formed within the Arctic basin has not increased. Zhang et al. (2004) and Simmonds and Keay (2009) also report an increase in the intensity of cyclones entering the Arctic from the mid-latitudes. Zhang et al. (2004) further found out that Arctic cyclone activity displays significant low-frequency variability, with a negative phase in the 1960s and a positive phase in the 1990s. Over smaller sea areas, such as the Bering and Chukchi Seas, the trends in cyclone activity since 1948 have been weak (Mesquita et al., 2010).

Since a strong storm event in the Beaufort Sea during August 2012 (Simmonds and Rudeva, 2012), the effect of summer storms on sea ice has received a lot of attention. According to a modelling study by Zhang et al. (2013), the strong melt was largely due to a quadrupling in bottom melt, caused by storm-driven enhanced mixing in the ocean boundary layer. Zhang et al. (2013) argued, however, that a record minimum ice extent would have been reached in 2012 even without the storm. It should be noted that summer cyclones in the Arctic are climatologically weak and usually not generating storm-force winds (defined as 10-minute mean wind speed exceeding 20 m s-1). For example, the SHEBA and Tara ice stations did not experience a single summer day with wind speed exceeding 20 m s^{-1} (Vihma et al., 2008). According to Walsh et al. (2011), storm activity has increased at some locations in the North American Arctic, but there are no indications of systematic increases in storminess in the Arctic over the past half century, and no significant trend over the central Arctic in storm intensity can be found.

Evaluating published results, a problem in climatological cyclone analyses is that it is difficult to fully distinguish between true and apparent changes in the cyclone occurrence and properties. Most studies rely on reanalysis data sets. The apparent changes may originate from changes in the amount, type and quality of observations assimilated into reanalyses. Above all, the number of high-latitude radiosonde sounding stations has decreased meanwhile the amount of satellite data has strongly increased. The results are also sensitive to the cyclone detection method applied (Neu et al., 2013). Several studies applying different reanalyses and cyclone detection methods however, have suggested an increase in the Arctic cyclone activity. It is potentially partly related to the sea ice decline, as the horizontal temperature gradient at the sea ice edge favours baroclinic instability, but interaction with lower latitudes cannot be ignored (Zhang et al., 2004; Trigo, 2006). On the basis of climate model experiments Solomon (2006) concluded that warmer climate with more water vapour in the atmosphere should yield stronger extratropical cyclones. According to Bengtsson et al. (2006; 2009), however, the number of cyclones in the Arctic does not necessarily depend on the changes in greenhouse gas concentrations. Another challenge in evaluating the results is related to the terminology used. Some authors write about cyclones while others write about storms, and the criteria used (among others, the lower threshold of wind speed for a system to be called storm) are often not mentioned. Given those uncertainties, results on cyclone climate in the Arctic needs to be taken with care. Research is necessary on the impact of the mentioned analysis problems on resulting cyclone frequencies and intensities.

4.2 Atmospheric transports of heat, moisture and aerosols

Anomalous large-scale transports of atmospheric moisture have been shown to contribute to rapid sea ice melt events such as the 2007 record low sea ice extent. Increased air specific humidity and, above all, cloud cover, enhanced long wave downward radiation (Graversen et al., 2011), which supports melting of sea ice.

On a more general level, Atmospheric transport of moist static energy from lower latitudes is the primary source of heat for the Arctic energy budget. Depending on the season this heat transport across 70°N is equivalent to 60-120 W/m² if evenly distributed over the polar cap (Nakamura and Oort, 1988; Serreze et al., 2007a; Skific and Francis, 2013; Semmler et al., 2005; Serreze and Barry, 2005;), which is weakest during April-May. On the annual average, the lateral heat transport exceeds the downward solar radiation. In the mass transport, the essential components are the air moisture, clouds, and aerosols. The transport of latent heat is equivalent to 10-25 W/m² (Serreze et al., 2007b). An indirect heating effect of moisture transport, via cloud formation and associated radiative effects, is, however, much larger (see section 4.3). Atmospheric heat transport has a strong effect, among others, on the inter-annual variability of the winter ice edge in the Bering and Barents Seas, the areas where the ice edge has the most freedom to vary. Francis and Hunter (2007) showed that from 1979 to 2005 the Bering Sea ice edge was controlled mainly by anomalies in easterly winds associated with the Aleutian Low, whereas the Barents Sea ice edge was affected by anomalies in southerly wind, in addition to a major influence of SST.

The transports of heat and moisture consist of the contributions by the background hemispheric circulation and by transient eddies. As an important part of the latter, synoptic-scale cyclones are responsible for most of the transport to the Arctic (e.g. Zhang et al., 2004). According to Jacobson and

Vihma (2010) transient cyclones contribute 80-90% of the total meridional moisture flux. The main moisture flux into the Arctic occurs in the Norwegian Sea and Bering Strait sectors and the main moisture export in the Canadian sector. The inter-annual variability in moisture transport is mainly driven by variability in cyclone activity over the Greenland Sea and East Siberian Sea (Sorteberg and Walsh, 2008).

Considerable uncertainty remains in the vertical distribution of moisture transport. According to rawinsonde sounding data, the meridional moisture flux across 70°N peaks approximately at the 850 hPa level (Overland and Turet, 1994, Serreze et al., 1995), whereas according to ERA-40 reanalysis the median peak level is in winter at the 930 hPa level and in other seasons at the 970–990 hPa level (Jakobson and Vihma, 2010).

In addition to heat and moisture, large-scale atmospheric transport is the main contribution to the concentration and composition of cloud-condensation nuclei. This is the case especially in winter (Garrett and Zhao, 2006). In summer over sea ice, aerosol concentrations in the boundary layer are generally low, but transport from lower latitudes may occur at higher elevations (Kupiszewski et al., 2013).

In general, not much is reported about trends in heat and moisture transport, although the effect of large transports on the September 2007 sea ice minimum has received attention (Graversen et al., 2011). The trends reported are very sensitive to the time period chosen. The ERA-40 reanalysis does not show any significant trend in the atmospheric moisture flux convergence over the Arctic Ocean during 1979–2001 (Serreze et al., 2006). Using satellite-based air temperatures and reanalysis products, Yang et al. (2010) detected periods of decreased and increased energy flux convergence in the Arctic: 25% of the cooling during a decade centered in the late eighties was due to decreasing poleward energy transport, and half of the warming during a decade centered in the late nineties was due to increasing poleward energy transport. Zhang et al. (2012) concluded that in the period of 1948-2008 the net atmospheric moisture transport to the Arctic has increased by 2.6% per decade. Model experiments have suggested increasing poleward transports in a warmer climate. On the basis of sensitivity tests on the surface energy budget, Lu and Cai (2009) suggested an enhancement of poleward moist static energy transport, and Solomon (2006) found that stronger extra-tropical cyclones (Section 4.1) yield increased northward heat and moisture transports.

Horizontal heat and moisture transports are affecting the sea ice cover via the radiative and turbulent heat fluxes. On the basis of ERA-Interim reanalysis, Maksimovich and Vihma (2012) calculated that an early melt onset in spring is favoured by large downward longwave radiation. This is typically associated with advection of warm and cloudy marine air masses from lower latitudes to the Arctic. Kapsch et al. (2013) report that in years with an end-of-summer sea-ice extent well below normal, a significantly enhanced transport of humid air is evident during the spring before, directed into the region where the ice retreat occurs. This enhanced transport of humid air leads to an anomalous convergence of humidity and to an increased cloudiness, connected to increased long wave downward radiation. Accordingly, the downwelling short-wave radiation is not decisive for the initiation of the melt, but acts as an amplifying factor later in the summer.

A further link between lower latitudes and Arctic climate change is seen in the atmospheric transport of sulphate aerosols (originating from burning of coal and oil) and black carbon (originating from combustion of diesel and biofuels) from anthropogenic sources into the Arctic. While sulphate aerosols are found to cool the atmosphere and surface due the increased net albedo, black carbon warms the air because of its increased absorption of solar radiation. Black carbon deposition on snow and ice may support melting due its reduced albedo (Section 3.5). During the past three decades, inflow of the cooling sulphate aerosols was reduced (Sharma et al., 2013), in contrast with an increased inflow of the warming black carbon (Serreze and Barrett, 2008). Shindell and Faluvegi

(2009) estimate an aerosol contribution of $1.09 \pm 0.81\ ^\circ$C to the Arctic surface temperature increase between 1976–2007, based on a reconstruction of aerosol radiative forcing. Thus, an influence of those processes to Arctic warming appears likely, although uncertainties exist, concerning compensating effects and emissions of both warming and cooling aerosols.

Assessing the reported findings, the seasonal and large-scale spatial variability in the transports of heat and moisture are reasonably well known. Also, consistent results exist relating humidity transports to sea ice melt. Reliable detection of trends is, however, very difficult, because of (a) large inter-annual and decadal variability, and (b) inaccuracy of reanalyses, both due to model deficiencies and decadal differences in the amount of observations available. Considerable uncertainty remains, among others, in the vertical distribution of moisture transport. There are also large differences between reanalyses in the accuracy of the closure of the atmospheric moisture budget.

4.3 Clouds, precipitation and evaporation

Clouds occur in the Arctic due to local condensation and lateral advection from lower latitudes. The strong effect of clouds on the Arctic sea ice heat budget is reported in several studies (Francis et al., 2005; Francis and Hunter, 2007; Stroeve et al., 2007; Schweiger et al., 2008a,b; Lu and Cai, 2009; Graversen and Wang, 2009; Graversen et al., 2011). For most of the year the cloud radiative forcing is positive, i.e. clouds increase the downward longwave radiation more than they reduce the downward shortwave radiation. In winter clouds may increase the downward longwave radiation by up to 90 $\mathrm{Wm^{-2}}$ (Overland and Guest, 1991; Minnet, 1999). On the basis of Russian drifting station data from 1968-1991, clouds significantly decrease the surface net radiation only in May – July (Chapman and Walsh, 1998), and on the basis of SHEBA data only in mid-summer (Intrieri et al. 2002; Shupe and Intrieri 2004). The representativeness of these observations for the present Arctic climate is, however, uncertain, because the cloud effect on net radiation is very sensitive to surface albedo, latitude, and cloud properties (Sedlar et al. 2011). The climatology of clouds and their properties are poorly known over the Arctic Ocean (better known for circum-Arctic observatories (Shupe 2011)). The radiative effects of clouds are very sensitive to the distribution of condensate content between liquid water and ice, warm liquid water clouds being much more effective in emitting longwave radiation (Shupe and Intrieri, 2004). The reanalyses-based results of Maksimovich and Vihma (2012) and Kapsch et al. (2013) (Section 4.2) are in accordance with SHEBA data, suggesting that the cloud forcing on net radiation over the Arctic sea ice is still positive in spring and early summer, when the snow melt on sea ice starts.

Excessive cloud cover in spring contributed to the September 2007 sea ice minimum (Graversen et al., 2011) whereas conclusions scatter on the effects of the anomalously clear skies from June through August 2007, which resulted in increased downwelling shortwave radiation; according to Kay et al. (2008) it was a major factor in the Beaufort Sea, and according to Schweiger et al. (2008b) it did not substantially contribute to the sea ice minimum, based on observations in the Chukchi Sea.

Changes in the cloud cover in the marine Arctic are not well known. Vihma et al. (2008) observed that the atmospheric transmissivity to shortwave radiation was significantly smaller during the Tara drift in April-September 2007 compared to Russian drifting stations in 1968-1990, which suggest an increase in cloud cover or optical thickness. Mostly on the basis of satellite data, Kay and Gettelman (2009) concluded that low cloud cover in early autumn has increased as a response to sea ice loss, but summer cloud cover does not depend on sea ice cover because of thermal decoupling. Increase in autumn cloud cover was detected also by Francis et al. (2009) and Palm et al. (2010). On the basis of synoptic observations reported from weather stations on land, drifting stations on sea ice, and ships, Eastman and Warren (2010) detected small positive pan-Arctic cloud cover trends in all seasons during the 1971-2009 period. Low clouds were primarily responsible for these trends. Focusing to the sea ice zone, clouds showed a tendency to increase with increasing air temperature and decreasing sea ice in

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all seasons except summer. Particularly in autumn, there was an increase in low clouds consistent with reduced sea ice, indicating that recent cloud changes may be enhancing the warming of the Arctic and accelerating the decline of sea ice (Eastman and Warren, 2010). On the basis of TOVS satellite data, however, Schweiger et al. (2008a) observed that sea ice retreat is linked to a *decrease* in low-level cloud amount and a simultaneous increase in mid-level clouds. The results on increasing cloud cover are consistent with the ensembles of 21st century projections by Vavrus et al. (2010), who found that clouds increased in autumn during periods of rapid sea ice loss.

It is noteworthy that ERA-Interim reanalyses yields different cloud cover trends than observations: spring is the only season with significant trends in Arctic average cloudiness; these trends are negative (Screen and Simmonds, 2010b). In general, the largest uncertainty and differences between different reanalysis data sets are related to depiction of clouds (Bromwich et al., 2007). Considering model experiments, Barton and Veron (2012) found that in the regional atmosphere model Polar WRF a low sea ice extent resulted in more clouds with larger liquid water paths.

It is difficult to quantify to what extent increases in air specific and relative humidity and cloud cover are due to sea ice decline or increased transports from lower latitudes. Recent studies have suggested increasing trends in the air moisture in the Arctic (Dee et al., 2011; Screen and Simmonds, 2010a,b; Rinke et al., 2009; Serreze et al., 2012). On the basis of three reanalyses (ERA-Interim, NASA-MERRA, and NCEP-CFSR) Serreze et al. (2012) have detected significant increasing trends in vertically integrated water vapour content in the period 1979-2010, in particular in the regions where the sea ice cover has decreased most and SST has increased most. Boisvert et al. (2013) studied evaporation from the Arctic Ocean and adjacent seas applying a new method (Boisvert et al., 2012): the air specific humidity was based on satellite data (Atmospheric Infrared Sounder onboard EOS Aqua satellite) and the wind speed on ERA-Interim reanalysis. Statistically significant seasonal decreasing trends in evaporation were found for December, January and February because of the dominating effect of increase in 2m air specific humidity, reducing the surface-air specific humidity difference in the Kara/Barents Seas, E. Greenland Sea and Baffin Bay regions, where there is some open water year round. Simultaneously the evaporation has slightly increased in the central Arctic, due to decreased sea ice concentration. The results of Boisvert et al. (2013) included similarities and differences with those of Screen and Simmonds (2010a), based on in-situ observations and ERA-Interim reanalysis. Screen and Simmonds (2010a) concluded on general increases in evaporation over the Arctic, but their study area did not include the Barents Sea and study period did not include November and December, which according to Boisvert et al. (2013) probably was the main reason for the different general trends.

Precipitation observation over Arctic land areas suggest that recent pan-Arctic precipitation exceeds the mean of 1950s by about 5%, and the years since 2000 have been wet both in terms of precipitation and river discharge (Walsh et al., 2011). According to Zhang et al. (2012), the Eurasian Arctic river discharge has increased by 1.8% per decade. This has accelerated in the latest decade and an unprecedented, record high discharge occurred in 2007 (Shiklomanov et al., 2009). The increasing trend has been attributed to warming effects, including intensifying precipitation minus evaporation, thawing permafrost, increasing greenness and reduced plant transpiration, but the causal physical processes have remained unclear (Zhang et al., 2012). These results are, however, for Arctic land areas; information on temporal changes over the Arctic Ocean is almost entirely based on atmospheric reanalyses. Contrary to pan-Arctic land areas, on the basis of ERA-Interim, Screen and Simmonds (2012) detected a decrease of total precipitation over the Arctic Ocean and Canadian Archipelago in 1989-2009. From the point of view of sea ice, however, it was more important that the summer snowfall had decreased by 40% and the rain had increased with a strong contribution to the recent decline (Section 3.5). Screen and Simmonds (2012) concluded that the decline in summer snowfall has likely contributed to the thinning of sea ice over recent decades. Contrary to findings by Screen and

Simmonds, experiments with a single regional atmosphere model by Porter et al. (2011) suggested that Arctic sea ice loss increases cloud cover, precipitation and evaporation in the Arctic.

In summary, clouds, precipitation, and evaporation are major factors in affecting the state and change of the Arctic climate system, but large problems remain. First, a major problem in evaluating the changes is that there are very few surface-based observations on clouds, precipitation and evaporation over the Arctic sea ice zone. Further, most cloud observations are qualitative, addressing the cloud coverage, levels, and types, which is not enough to estimate the radiative effects of clouds. Second, presentation of Arctic cloud physics, particularly for mixed-phase clouds, in reanalyses and climate models is liable to large errors and uncertainties (e.g. Tjernström et al., 2008; Morrison et al., 2012).

4.4 Vertical profile of Arctic warming

Different results have been presented on the vertical structure of warming in the Arctic atmosphere. On the basis of the ECMWF ERA-40 reanalysis for 1979-2001, Graversen et al. (2008) detected the maximum warming well above the Earth surface. They also found that in the summer half-year a significant part of the vertical structure of warming is explained by an increase in the atmospheric energy transport from lower latitudes to the Arctic. On the basis of the ERA-Interim reanalysis for 1989-2008, Screen and Simmonds (2010b) found, however, that the maximum Arctic warming has occurred at the Earth surface, decreasing with height in all seasons except summer. They further suggested that decreases in sea ice and snow cover have been the dominating causes of the Arctic amplification. The different results of Graversen et al. (2008) and Screen and Simmonds (2010b) were related to different time periods, studied on the basis of different reanalyses. Later, on the basis of climate model experiments, Screen et al. (2012) suggested that local changes in sea ice concentration and SST explain a large portion of the observed Arctic near-surface warming, whereas the majority of observed warming aloft is related to remote SST changes, which have contributed to heating of the air-masses that are transported from lower latitudes to the Arctic. According to Screen et al. (2012), the direct radiative forcing due to observed changes in greenhouse gases, ozone, aerosols, and solar output has primarily contributed to Arctic tropospheric warming in summer.

Analyses of the vertical profile of Arctic warming are liable to uncertainties. Recent studies have shown that in the central Arctic reanalyses have large errors in near-surface variables (Lüpkes et al., 2010; Jakobson et al., 2012) and large mutual differences in the vertical structure at least up to the mid-troposphere (Chung et al., 2013). Possibilities to use other means to study the vertical profile of Arctic warming are, however, limited. In-situ observations over the Arctic Ocean are mostly restricted to the lowest tens of metres (buoys, ships). Radiosonde and tethersonde soundings have been made at ships and drifting ice stations, but most of these observations cover short periods only. An exception is the long-lasting radiosonde sounding program at the Russian ice stations from 1954 to 1991 (and to some extent also since 2003). The Russian drifting station data have been an important basis for climatology of the vertical air temperature structure (e.g. Serreze et al., 1992) and, combined with shorter periods of data from more recent years, could be more systematically utilized to study the vertical structure of warming over the Arctic Ocean. Only a few studies of this kind have been carried out so far. Vihma et al. (2008) showed that, compared to the mean conditions in the Russian stations, summer 2007 was clearly warmer and moister at the altitudes from 200 to 1000 m, although the July mean 2-m temperature had not increased at all. As long as the surface temperature is restricted by the melting point, the near-surface air temperatures over inner parts of large ice-covered areas cannot raise much above the melting point.

Satellite and surface (ship/ice/land) based remote sensing methods have a potential to provide better understanding of the vertical profile of air temperature trends over the Arctic Ocean. The time series of high-quality data are getting long enough to yield interesting results about inter-annual variations. For example, the Atmospheric Infrared Sounder has operated since 2003, and Devasthale et al. (2010)

found that summer 2007 was 1.5 to 3.0 K warmer than the mean of 2003-2006 and 2008 in a thick layer from the surface up to the 400 hPa level.

Despite the dominating warming trends, also periodic cooling trends have been detected in the Arctic. Focusing on the 1998–2011 period, Chung et al. (2013) demonstrated that four reanalyses products (ERA-Interim, CFSR, MERRA and NCEP II) show a cooling trend in the Arctic-mean 500 hPa temperature in autumn, and this is supported by coastal rawinsonde sounding data. No signs on recent near-surface cooling have been observed over the Arctic Ocean, but a widespread near-surface winter cooling has been observed over land areas in northern Eurasia and eastern North America since approximately 1988 (Cohen et al., 2012).

The ABL thickness, controlling the ABL heat capacity, is an important factor affecting the vertical structure of temperature trends in conditions of both warming and cooling. In the Arctic the shallower ABL, with a heat capacity smaller than at lower latitudes, is a factor contributing to the Arctic amplification (see section 2). It may also partly explain the fact that the Arctic warming has been larger in winter than summer (e.g. Walsh et al., 2011) and that global warming has been larger during night than daytime (Graversen and Wang, 2009; Esau et al., 2012). The stronger near-surface cooling of the Arctic compared to global temperatures during 1940-1970 (Chylek et al., 2009) may also have been affected by the smaller heat capacity of the thin ABL in the Arctic.

Studies on the vertical profile of Arctic climate change benefit from recent advances in understanding the mechanisms of stratosphere-troposphere coupling. It has been known for over a decade that a cold anomaly in the stratosphere typically results in a positive phase of AO and NAO (Wallace, 2000; Baldwin and Dunkerton, 2001; Karpechko and Manzini, 2012), and stratospheric circulation influences the vertical wind shear near the tropopause, and so the baroclinic instability across the depth of the troposphere, which affects the formation and growth of cyclones (Wittman et al. 2004). Recent advances in the field include studies that demonstrate how disturbances in the Earth surface, e.g. snow cover, generate vertically propagating planetary waves which reach the stratosphere and then have a lagged downward influence on the near-surface weather and climate (Cohen et al., 2007; Orsolini and Kvamstö, 2009; Allen and Zender, 2011; Peings and Magnusdottir, 2013). Bitz and Polvani (2012) found that the effect of stratospheric ozone depletion is to warm the surface and the ocean to a depth of 1000 m and to significantly reduce the sea ice extent.

5. Recent advancement of understanding of the role of the ocean for sea ice changes.

The ocean's role in the Arctic climate system is the least explored, due to even more difficult accessibility compared to the atmosphere and sea ice. Mooring-based observations and ship-based expeditions during IPY as well as Ice-Tethered Platforms (ITPs) and first Automatic Underwater Vehicles (AUVs) have started to improve the situation, together with numerical process studies and climate change simulations.

The general picture of Arctic Ocean hydrology and circulations includes a shallow surface layer of relatively fresh and cold water dominated by river runoff. That upper Polar Surface Water is largely isolating sea ice from the underlying warmer cores of salty Atlantic water between 300 and 500 meters and relatively fresh Pacific water between 40 and 80 meters depth (Bourgain and Gascard, 2012). The latter is largely limited to the Canadian Basin and adjacent seas.

In this section, we review recent progress in understanding the role of warm ocean inflow for sea ice change in conjunction with the ocean's part in ocean-sea ice-atmosphere feedbacks. While changes in ocean temperature and circulation are obvious, it appears more difficult to establish a link to sea ice changes.

5.1 Transports and pathways of water

The passages connecting the Arctic Ocean with the world ocean measure just several tens to hundreds km in case of the Fram Strait, Bering Strait and Canadian Archipelago. The Barents Sea opening with its 1000 km scale is the exception. Pacific water enters the Arctic through the Bering Strait. The basic reason for the flow direction is a higher steric sea level in the Pacific compared to the Atlantic, giving rise to a wide trans-Arctic drift from the Bering Strait to Fram Strait. In the Atlantic sector, the Canadian Archipelago is an export gateway for water volume and for freshwater (Rudels, 2011). Fram Strait features southward transport of freshwater, salt and sea ice. The Canadian Archipelago carries about 50% the freshwater transport of the Fram Strait (Dickson et al. 2007). Both the Fram Strait and Barents opening experience northward transport of Atlantic water of equal magnitude. Recent high-resolution numerical flow simulations point to a volume inflow into the Arctic equally divided, but the heat entering the Arctic Ocean largely through the Fram Strait (Aksenov et al. 2010).

Pathways of northward ocean transports into the Fram Strait and Barents Sea opening are rather complex. Here we focus on the fate of the Atlantic water within the Arctic Ocean and its potential to impact sea ice. As a long known general feature of Fram Strait flow, the East Greenland current flows southward while the West Spitsbergen Current (WSC) penetrates into the Arctic Ocean. That Atlantic water returns in parts (2 Sv) due to a local recirculation (Aagaard and Greisman, 1975; Marnela et al., 2012). The remaining part, ca 2-4 Sv (Schauer et al. 2008, and Beszcynska-Möller et al., 2012) of the WSC proceeds eastwards along the continental slope in two different branches (Schauer et al. 2004). Little is known about its further processing by turbulent eddies. Here we rely on high-resolution numerical models. Aksenov et al (2010), using a numerical model of 1/12° horizontal resolution, find that after passing the Fram Strait, the Atlantic water inflow splits into a deeper and a shallower branch following the shelf break of Svalbard, and then reuniting east of the Yermak Plateau into a single Fram Strait branch.

An overall increase in northward flowing Fram-Strait temperature and transports was found after 1999 and 2004 (Schauer et al. 2004, Dmitrenko et al. 2008, Beszcynska-Möller et al., 2012). Indication for increased inflow was seen already in the early 1990's when Atlantic Water was observed in the southern Makarov Basin which before was dominated by Pacific waters (McLaughlin et al., 1996; Smith et al., 1999). Multi-year pulse-like anomalies which formed in the North Atlantic and Nordic Seas have been observed passing Fram Strait and further propagating eastwards along the Arctic continental slope. Mooring-based observations in Fram Strait and oceanographic surveys during the DAMOCLES project and earlier give an overall warming trend in the northward flowing Atlantic water of 0.06°C per year, between 1997 and 2010 (Beszcynska-Möller et al., 2012), although the actual warming trend in the northward WSC ceased after 2007, but still is elevated compared to the early 1990's. (Polyakov 2011). On a longer time scale, proxy data from marine sediments off Western Svalbard (79°N) reveals that recent Atlantic water temperatures are unprecedented compared to the past 2000 years (Spielhagen et al., 2011). The volume transport variability in the WSC is limited to the offshore branch west of the Spitsbergen shelf, and no statistically significant trend can be found in the Artic Water (AW) volume transport.

5.2 Northward heat transport

Signals of increasing northward heat transport before 2007 can be traced along the Siberian shelf (Polyakov et al. 2008, 2011, Bourgain and Gascard, 2012) all the way to the Laptev slope (after 4.5 – 5 years), Chukchi shelf and even at the Lomonosov Ridge and in the Makarov Basin (Rudels et al. 2012).

In the Eurasian and Makarov Basins, AW warming of up to 1ºC was observed in 2007 relative to the 1990s average (Polyakov et al. 2010). At the same time, the upper AW layers were raised by up to 75-90 m in the central Arctic Ocean, related to a weakening of the Eurasian Basin upper-ocean stratification (Polyakov et al. 2010).

Even a seasonal cycle, originating from the AW inflow at Fram Strait, has been found to survive mixing processes and transformation into Arctic intermediate water (Ivanov et al. 2009). Integrated views based on mooring observations and high-resolution ocean models (Lique and Steele, 2012) show that the AW seasonal cycle signal is advected from the Fram Strait up to the St. Anna Trough and then re-energized by the Barents Sea Branch. The seasonal AW temperature signal survives within the Nansen basin. Interannual changes in the seasonal cycle amplitude can be as large as the mean seasonal cycle amplitude.

The observed interannual warming of AW in the Arctic Ocean implies pools of anomalously low density. These are expected to slowly drain back south into the Nordic Seas (Karcher et al. 2011), with the anticipated effect of a reduced Denmark Strait overflow into the North Atlantic Ocean.

While ample progress has been made concerning the monitoring of the AW inflow signal and understanding of its fate, the more difficult task of understanding the impact on sea ice coverage has just started to give results. It is hypothesized that the changes in the Eurasian Basin (warming and up-lifting of the AW layer) facilitated greater upward transfer of AW heat to the ocean surface layer, thus impacting ice melt (Polyakov et al. 2010).

5.3 Links between ocean heat transport and sea ice melt

Ocean heat transport into the Arctic is linked to the Atlantic multidecadal oscillation (AMO), both in observations (Chylek et al.,2009; Wood and Overland, 2010) and in climate model studies (e.g. Semenov, 2008). However, there is also an indication of increasing heat transport despite a recently reduced AMO (Koenigk and Brodeau, 2012). A general large scale relationship between ocean northward heat transport in the Norwegian Sea and Arctic ice cover is considered to be well established (e.g. Sandø et al. 2010, Smedsrud et al. 2010).

It was long unclear to what extent processes connecting Atlantic water with ice melt can be described realistically. Despite strong surface cooling of inflowing Atlantic water into the Barents Sea, those waters have warmed during the last 30 years by 0.3ºC averaged over the Barents Sea. (Levitus et al. 2009). Recent findings in the area are often based on lengthening of pre-existing time series eventually enabling new conclusions. Already Vinje (2001) found that observed temperature anomalies in the central Norwegian Sea are significantly correlated with the Barents Sea sea ice extent with a lag of two years. Later, according to Årthun et al. (2012), observed sea ice reduction in the Barents sea (up to 50% on annual mean between 1998 and 2008) has occurred concurrent with an increase in observed Atlantic heat transport due to both strengthening and warming of the inflow. The winter mean ice extent between 1979 and 1997 is clearly affected by the inflowing warm AW, with an ice margin shifted towards the north and east (Årthun and Schrum, 2010).

Observation-based heat budget calculations by Årthun et al. (2012) show that Barents Sea heat content, ocean-atmosphere heat fluxes and sea ice cover in the Barents sea respond on a monthly to annual time scale to increased heat transports from the Norwegian Sea. Barents Sea sea ice bottom heat uptake from the ocean is proportional to the water temperature (Rudels et al., 1999), and thus should have increased during the Barents Sea warming. On annual average however, the ice bottom experiences freezing while net melting occurs at the top. The Barents sea ice cover is rather reduced by the warming waters capability to prevent freezing due to a longer period of cooling down water to

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the freezing point, especially in the central and eastern Barents Sea. Those relationships and lags are confirmed by a local ocean-sea ice circulation model (Årthun et al. 2012).

Also coupled climate models often show a relation between northward ocean heat transport from the Nordic Seas into the Arctic Ocean and the Arctic sea ice cover. Holland et al. (2006) find pulse-like increases in ocean heat transport leading ice melt events by a lag of 1-2 years, showing that rapid increases in heat transport can trigger ice melt events in models. Koenigk et al. (2011) find ice thickness to be highly negatively correlated with the ocean meridional overturning circulation (MOC) due to larger than normal ocean heat transport to the north during periods of anomalously strong MOC. Bitz et al. (2006) even show positive heat transport events independent of the MOC. In such cases, the ocean heat transport events represent a positive feedback responding to reduced sea ice, increased brine release, strengthening convection and in turn bolstering the inflow of warm Atlantic water (Bitz et al., 2006). Koenigk et al (2011) using a regional coupled climate model, find that enhanced surface heating in the Nordic Seas or North Atlantic contributes to increasing northward ocean heat transports in a future climate change projection.

Recent results based on an ensemble of future climate projections (with the EC-Earth GCM) suggests that heat transport through the Barents Sea opening governs sea ice variations in the Barents and Kara Sea on decadal time scales. Koenigk et al. (2012) indicate that the increasing ocean heat transport strongly contributes to the reduced sea ice cover in the Barents and Kara Sea region and thus hypothetically also contributes to the Arctic temperature amplification of the global climate warming (see section 2.1). About 50% of the inflowing ocean heat anomaly in the 21st century scenario ensemble is either used to melt sea ice or is passed to the atmosphere north of 70^{o} North.

Intense water mass transformation of the Atlantic inflow occurs not only in the Barents Sea, but also in the Kara Sea and Nansen Basin through atmospheric-ocean heat-exchange and ice edge processes (Årthun and Schrum 2010). Recent observations point to interaction processes along the shelf break north of Spitsbergen and in the Barents and Kara Seas. In this area, the Atlantic water has the strongest potential to affect the sea ice. Temperature/Salinity (T/S) profiles at the Barents Sea shelf break are lacking a summer sub-surface temperature minimum between the warm summer surface and the warm Atlantic water layer. The Barents Sea shelf area is unique in the Arctic for such conditions. This means that at this location, upward heat flow from the Atlantic water layer to the surface and the ice is likely (Rudels et al. 2012). The reasons behind this phenomenon are likely more intense vertical homogenization during winter, including deeper layers of Atlantic water. Rudels et al. (2012) relate the homogenzation to mechanical mixing processes due to wind and the topographic slope which might increase the entrainment of Atlantic water into the surface layer.

5.4 Pacific water inflow and sea ice melt.

The inflow of Pacific water through the Bering Strait is traditionally estimated at about 0.8 Sv. (e.g. Coachman and Aagaard, 1988; Aagaard and Carmack, 1989) and confirmed later as the long-term annual mean (e.g. Woodgate and Aagaard, 2005). Strong seasonality in transport, temperature and salinity has been found (Woodgate and Aagaard, 2005).

Heat fluxes into the Arctic Ocean through the Bering Strait increased from 2001 to 2011 by a factor of 2 to a maximum of $5 * 10^{20}$ J/yr, with peaks in 2007 and 2011 (Woodgate et al. 2010, 2012). The difference of the annual heat fluxes between 2001 and 2007 could potentially melt $1.5 * 10^{6}$ km^2 of 1 m thick ice, corresponding to about 1/3 of the seasonal sea-ice loss during the 2007 summer event.

The warming signal originating from the Bering Strait, propagated into the interior of the Canadian basin during the mid and late 2000s, leading to a warming of the subsurface Pacific Summer Water between 1997 and 2008 (Bourgain and Gascard 2012). A temperature increase in the Pacific layer below 40 m depth can potentially promote summer melt and reduce winter growth. Pacific Summer Water has been proposed to initially trigger the onset of seasonal sea-ice bottom melt (Woodgate et al. 2010, 2012), and feeding a winter time subsurface temperature maximum under the ice (Toole et al., 2010). This might contribute to sea ice retreat in the western Arctic. However, little is known about the mechanisms that actually bring the heat in contact with the ice. Entrainment of the Pacific Summer water into the mixed layer has not been observed to our knowledge. Mixed layer studies rather indicate ongoing isolation of the Pacific Summer water from the mixed layer (Toole et al., 2010).

More well established is the role of the ocean in melting ice in response to local seasonal solar heating of the upper ocean. Summer insolation through leads and open water areas leads to increased surface temperature. Steele et al. (2008) find an upper ocean warming since the 1990s with a maximum temperature of 5^0C during summer 2007. Between 1979 and 2005, 89% of the Arctic Ocean surface area experienced an increase in the solar energy absorption of up to 5% per year (Perovich and Polashenski, 2012).

In the Canadian Basin, solar-driven surface temperature increase is quickly isolated by freshwater from melting sea ice, the heat remains located between 25 and 35 m. Contact with the surface can be re-established by wind induced vertical mixing, leading to melting at the ice edge and lead areas. Depending on the viability of the isolating freshwater layer, the sub-surface heat storage can contribute to winter ice melt or reduced winter ice freezing (Jackson et al., 2010).

6.Integrative summary and prospects

This article reviews recent progress in understanding of the decline of Arctic sea ice. Ice cover shows a shrinking trend at least since the 1970s, which is reflected in sea ice extent, thickness and volume. We are witnessing an Arctic sea ice pack which is thinning, becoming younger and more moveable, with a reducing albedo and lengthened melting season. All this makes the ice cover more susceptible for quick response to forcing from a warming earth system. Information on the mechanisms connected to the sea ice decline broadened during the 1990's and huge knowledge gains were possible due to intensified efforts after the year 2000 when the sea ice reduction accelerated. Major contributions were made from the International Polar year (IPY) and connected programs such as DAMOCLES and SEARCH and further initiatives. DAMOCLES studies on sea ice remote sensing are summarized in Heygster et al. (2012) and those on recent advance related to small-scale physical processes by Vihma et al. (2013).

The term “new Arctic” has been used to characterize a fundamental regime shift from predominantly multi-year ice to enhanced fractions of seasonal and generally thinner ice. Sea ice erodes both from the top and from the bottom, forced by atmospheric warming, changes in circulation and transports, as well as by increased ocean heat transports especially in the Barents Sea. In the Atlantic sector, the relation of large-scale ocean heat transport and sea ice extent is well established. Direct forcing of the sea ice decline by changing character of Pacific water inflow through the Bering Strait is unlikely to play a role. Instead, the increased rates of bottom melting in the Pacific sector can rather be related to increased leads and associated ocean mixing.

Sea ice thickness has clearly decreased since the 1970s from a winter mean estimate of 3.8 m down to 1.9 m in 2008. The relative decline of sea ice volume is even stronger due to simultaneous ice concentration reduction. Uncertainties of the sea ice volume trend estimates exist (about -875 ± 257 $km^3\ a^{-1}$ in winter) due to sparse direct observations and poorly bounded assumptions of parameters needed for satellite signal interpretation.

Arctic sea ice cover variability is both internally generated (within the Arctic) and externally forced (by varying hemisphere scale conditions). The relative importance of those influences varies in time and depends on the state of large-scale atmospheric circulation. Northerly wind anomalies in the Atlantic sector of the Arctic support ice export and favour external control of the Arctic variability (i.e. small internally generated variability), likely due to hemisphere scale influences on the wind anomalies, which are forcing the ice export. Internally generated sea ice variability is particularly large during periods when the ice volume increases.

Sea ice drift velocities have increased since the 1950's, partly due to increasing wind speeds and partly due to reduced sea ice strength. At least after 1989, inter-annual variability in ice drift speed appears to be connected to wind variability, while the trend in drift speed is rather related to ice thinning and the reducing mechanical strength, which are both associated with transformation of multi-year to first-year ice.

Record low summer sea ice extents after the year 2000 delivered additional information on relevant mechanisms for the ice decline. The event in September 2007 commenced with increased poleward ice drift, partly in the form of first-year ice. Anomalously high melt pond fractions were observed during the summers of 2007 and 2012, leading to reduced surface albedos. Increased convergence of meridional transport of moisture lead to reduced atmospheric short wave transmissivity, enhanced cloud cover and intensified long wave radiative melting during summer 2007. That event also highlighted dynamic effects of a changed atmospheric circulation with enhanced meridional transport components. Pronounced CAI and DA anomalies during summer 2007 were responsible for increased ice transport, while the 2012 event occurred under comparatively regular atmospheric conditions, except for an anomalously strong summer storm in August.

Additional influences on the sea ice decline originate from a pronounced decline in summer snowfall, which has been observed since the late 1980s. Generally enhanced transport of humid air is found in spring of those years where the end-of-summer sea-ice extent is well below normal. Other observations accompanied with the ice reduction are a longer melting period between melt onset in spring and the freeze-up in fall. Black carbon deposition on sea ice more efficiently absorbs radiation for young sea ice, which enables stronger melting on the growing area of one-year sea ice.

There are additional candidates potentially important for explaining the sea ice decline, but either no signal can be detected, or results are inconclusive. or contradicting. While a northward shift of cyclone activity is undisputed, the systematic changes in cyclone intensity remain unclear due to strong temporal variability and changes in the amount and quality of in-situ and remote sensing observations assimilated into atmospheric reanalyses. Further, comparison of individual studies is made difficult by differences in terminology used and methodology applied, among others in the cyclone tracking algorithms. Scientific opinions diverges on the possibility to draw conclusions from observations. There are no clear indications of systematic increases in storminess in the Arctic over the past half century. Although large both in 2007 and 2012, the fraction of melt ponds does not show a statistically significant trend during the last years and decade. Considerable uncertainty exists in the moisture transport into the Arctic (among others, in its vertical distribution), strongly affecting the cloud radiative forcing of the sea ice cover.

Arctic temperatures have risen to a level, which likely is unprecedented during the last 2000 years. The Arctic warming is enhanced by an Arctic amplification of the global warming signal, which is a result of the climate's internal response to changing radiative forcing. Arctic amplification is both supported by the sea ice reduction and is at the same time accelerating the ice decline. In addition to the long anticipated sea ice-albedo feedback, cloud and water vapour feedbacks, combined temperature feedback (lapse rate and Planck) and atmospheric circulation feedbacks play a role. The amplitude of the feedback depends on the state of the Arctic, its sea ice cover and planetary boundary layer stability. An emerging Arctic amplification of the global warming by e.g. the sea ice-albedo feedback can regionally activate and strengthen additional feedbacks such as the water vapour feedback with the result of an enhanced Arctic amplification. Consistently, increasing trends in vertically integrated water vapour content have been found particularly in the regions where the sea ice cover has decreased most and SST has increased most, leading to a locally enhanced tropospheric warming. According to observations, low cloud coverage has increased particularly in autumn but slightly also in other seasons. The reasons for the increase are, however, not clear. Sea ice decline itself favours evaporation but, according to Boisvert et al. (2013), winter evaporation in the marine Arctic has decreased in 2003-2011 and according to Schweiger et al. (2008a) sea ice decline is linked to a decrease in low cloud amount. Those apparent contradictions may be explained by the competing effects of decreased sea ice cover and increased advection of moist, cloudy air masses to the Arctic

Reasonable concepts explaining the Arctic amplification exist, although quantitative understanding is lacking. The different feedback mechanisms involved in the shaping of Arctic amplification depend on each other and partly compensate for each other in a self-adjusting way if single feedback types are suppressed. This suggests an Arctic amplification which is robust and not dependent on individual mechanisms.

Atmosphere, sea ice and ocean processes interact in non-linear ways on various scales under a global climate forcing. The Arctic sea ice extent shows a trend towards less ice, though is superimposed by oscillations reflecting the various influences. Each record low sea ice extent is followed by a partial recovery. Consulting climate change projections, even decadal scale periods of new low records can potentially alternate with periods of partially recovered sea ice (e.g. Massonet et al., 2012). The recent distinct recovery of summer sea ice extent in September 2013 might give a glimpse of the range of variability to be expected during the coming decades. It also illustrates a debate on possible tipping points for the sea ice cover.

Model studies of different complexity agree on a return of the sea ice cover under conditions of reducing climate forcing, e.g. reduced greenhouse gas concentrations (Tietsche et al., 2011; Stranne and Björk, 2012). In that sense, a tipping point of no immediate return does not exist. If the atmospheric forcing changes trends, sea ice can be re-established within just a few years. However, there is ample indication for a point of increased destabilization of the ice which justifies the term “New Arctic”. The decrease of extent, thickness and volume has distinctly accelerated around the year 2000. Positive feedbacks due to reduced sea ice and snow albedo are clearly detectable, often with stronger amplitude after the millennium shift. The accelerated development is further supported by the increasing prevalence of thinner and younger ice, which is more susceptible to further atmospheric warming and associated circulation changes, and even more sensitive to the albedo effects of soot deposition.

Climate prediction is an emerging science branch, still very much unexplored, but with well-founded hopes. Predictability studies with climate models indicate that sea ice anomalies can potentially persist for several years (Holland et al. 2011, Koenigk et al. 2009; Tietsche et al. 2013), a situation which allows for potential predictive skill of both sea ice and atmospheric conditions at least on a multi-year average. Potential predictability on multi-year time scales is high for the Arctic due to decadal scale ocean variability and due to signal storage capability in sea ice and ocean. Note that the potential predictability refers to climate conditions as simulated by climate models, typically under-representing the complexity of processes. Ongoing and upcoming projects (e.g. SPECS and the CMIP6 decadal prediction experiments) promise quick knowledge gains on the real-world potential. Current retro-active prediction experiments give good predictability for the Arctic area (Doblas-Reyes, 2013)

On the down side for predictability prospects is the thinning of the sea ice, which possibly reduces predictability due to lower signal storage capacity in the ice and increased interannual variability. It is unknown to what extent this can be compensated by greater amounts of heat anomalies stored in the ocean. Predictability in marginal ice areas such as the Labrador Sea and the Barents Sea are clearly influenced by largely predictable oscillations in ocean circulation and heat transports. A careful development of the future prospect of Arctic climate predictability requires accurate observation of Arctic ocean layers periodically in contact with the atmosphere, both for understanding storage processes and for a precise initialization of predictions. Furthermore, studies on future transformation between different atmospheric oscillation patterns (such as AO, DA and CAI in the past) will be essential for understanding the real potential of Arctic climate prediction.

For a proper exploration of climate prediction, it is essential to understand drivers and describe feedbacks of Arctic predictability. Studies such as reviewed here are therefore key, not only to describe Arctic climate change, but also for providing process understanding with the request to be properly reflected in prediction systems. A challenge in practical prediction efforts is an appropriate initialization of the ocean state including Arctic sea ice concentration, thickness and ocean temperature, which requires access to observations and exploration of initialization techniques. Also from that initialization point of view, observations of the state of the Arctic are essential.

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Figures

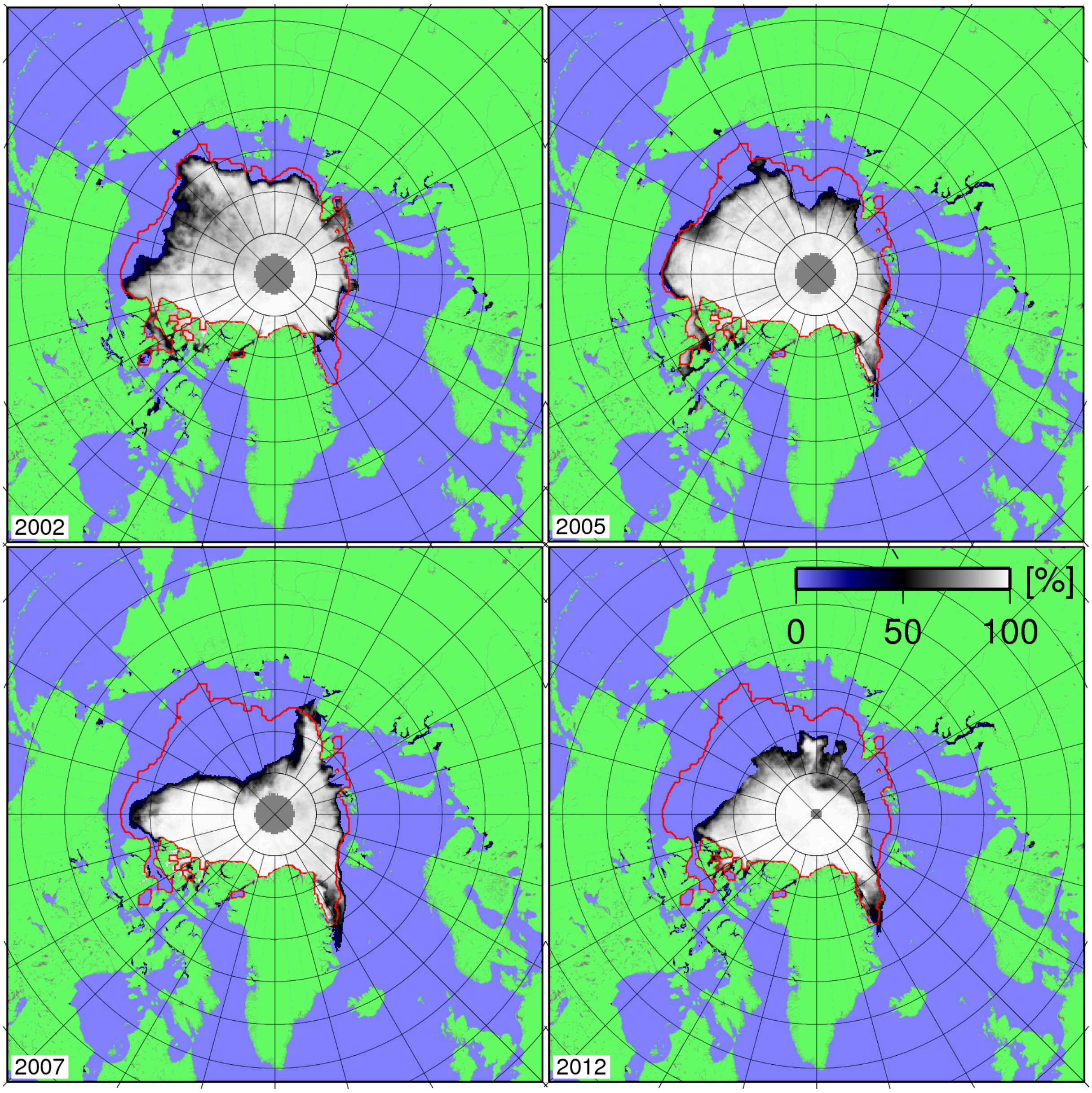

Figure 1: Monthly mean sea ice concentration (white to blue), based on SSM/I data for September 2002, 2005, 2007 and 2012, with the average ice margin (red) for the years 1992-2006. Pictures provided by Lars Kaleschke from University of Hamburg. The SSM/I algorithms are described by Kaleschke et al. (2001).

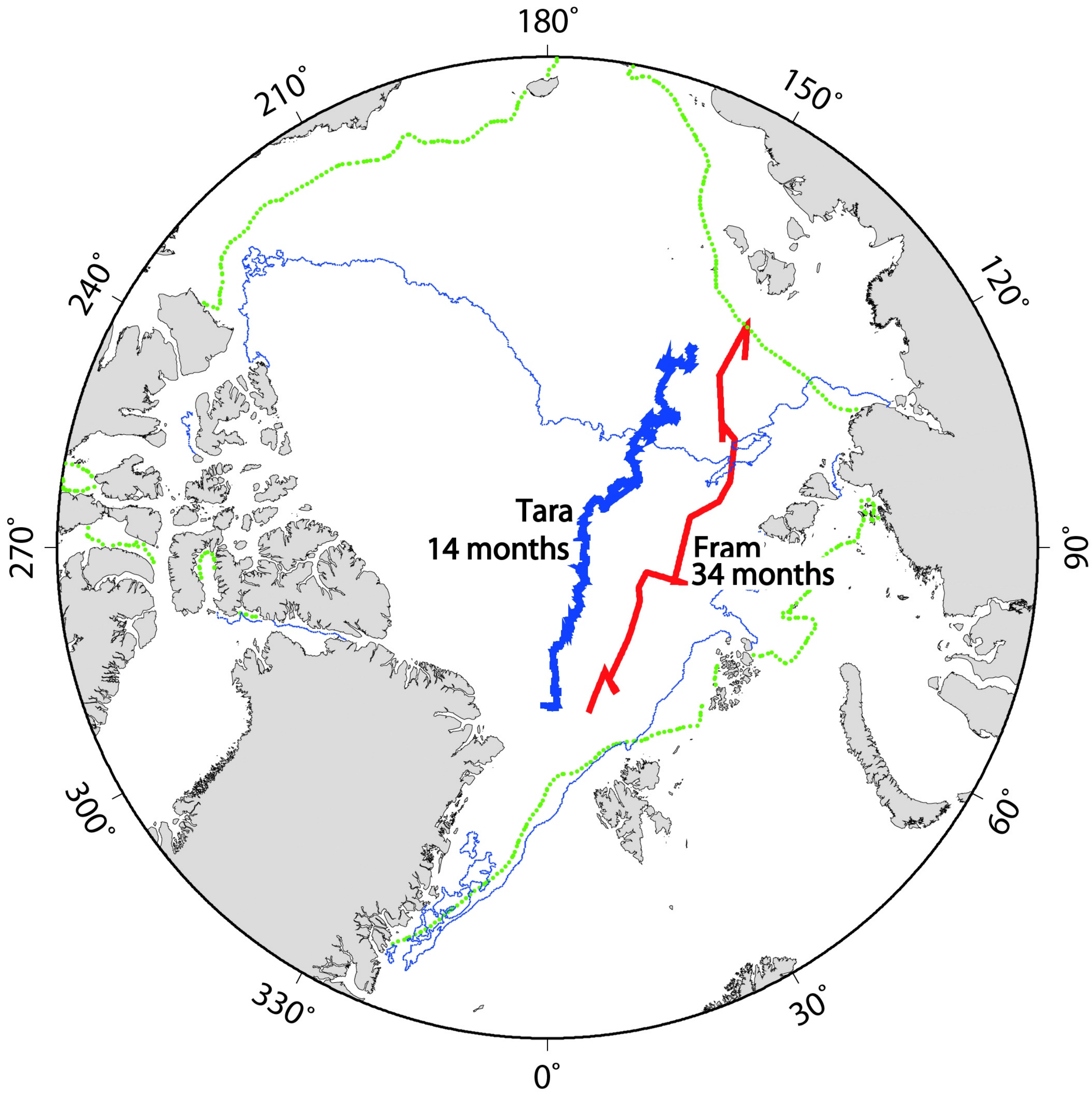

Figure 2: Drift trajectories of the vessels Tara (blue, November 2006 – January 2008) and Fram (red, October 1893 – August 1896). Further, the sea ice edges are displayed for September 2007 (blue) and for the September mean 1979–1983 (green).

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