Model-Data Fusion Studies of Pacific Arctic Climate and Ice-Ocean Processes

Jia Wang1, Hajo Eicken2, Yanling Yu3, X. Bai4, Jinlun Zhang3, H. Hu4, Moto Ikeda5, Kohei Mizobata6, and Jim Overland7

1 NOAA Great Lakes Environmental Research Laboratory (GLERL), 4840 South State Road, Ann Arbor, MI 48108. Tel: 734-741-2281; Email: Jia.Wang@noaa.gov

2 University of Alaska Fairbanks, Geophysical Institute, Fairbanks, AK 99775, hajo.eicken@gi.alaska.edu

3 University of Washington, Arctic Science Center, Seattle, WA. yy8@u.washington.edu, zhang@apl.washington.edu

4 Cooperative Institute for Limnology and Ecosystems Research (CILER), School of Natural Resources and Environment, University of Michigan, 4840 South State Road, Ann Arbor, MI 48108. Email: xuezhi.bai@noaa.gov, hghu@umich.edu

5 Hokkaido University, Graduate School of Earth and Environmental Sciences, Sapporo, Japan mikeda@ees.hokudai.ac.jp

6 Tokyo University of Marine Science and Technology, Department of Ocean Sciences, 4-5-7, Kounan, Minato-ku, Tokyo, 108-8477, Japan. Email: mizobata@kaiyodai.ac.jp

7 NOAA Pacific Environmental Research Laboratory, Seattle, WA.
James.E.Overland@noaa.gov

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Abstract

In this chapter, we investigate several emerging climate and climate-related dynamic and thermodynamic ice-ocean processes in the Pacific Arctic region (PAR) using both measurements and coupled ice-ocean models. These emerging scientific topics include 1) the Arctic Dipole Anomaly (DA) pattern dominating Arctic sea ice advection and reduction, and contributing substantially to a series of extreme summer sea ice extent minima; 2) the relationship between the DA and Bering Strait heat transport; 3) thermodynamic and dynamic controls of land fast ice along the Alaskan Beaufort coast; 4) the formation and maintenance of the Cold Pool on the northern Bering Shelf; and 5) the connection between the long-term summer sea ice reduction and ice-ocean albedo feedback, and the DA’s role in this feedback loop. Based on the data diagnosis and modeling, we examine these processes in depth in this chapter. During the early 1990s when the +AO dominated, the Arctic Ocean experienced a significant warming, resulting in thinning and shrinking of sea ice, due to significant heat transport from the northern North Atlantic (the Gulf Stream extension). Inside the Arctic Ocean, an anomalous cyclonic circulation and ice flow regime prevailed. However, since 1995, when the AO was in the neutral or negative territory, the +DA dominated the Arctic climate and ice-ocean circulation, leading to a series of changes in the Arctic. The most significant change was a series of record-breaking sea ice minima, reaching an all time low (since start of satellite observations in 1979) in the summer of 2007, in particular in the PAR. A significant meridional anomaly in the ocean and sea ice flow was substantial during the +DA, which accelerated the Transpolar Drift Stream (TDS) and caused anomalous meridional transport, in contrast with the anomalously cyclonic zonal flow regime in the early 1990s during the positive phase of the AO.
1. Introduction

The northern North Pacific Ocean, including the Bering Sea, is among the most productive marine ecosystems in the world, as evidenced by large populations of marine and freshwater fish, birds, and mammals. Fish and shellfish from these regions constitute more than 10% of total seafood harvest of the world and about 52% of the U.S. As a result, the productivity is critical not only to the U.S. economy, but also to the economy of surrounding countries.

The Bering Sea (Fig. 1) is a complex semi-enclosed sea with shallow shelves, shelf breaks, and deep basins. The ocean circulation pattern is complicated. The Alaskan Stream (AS) mainly flows along the Aleutian Peninsula and provides some of its water via Aleutian passes to the Bering Slope Current (BSC) and the Alaskan Coastal Water (ACW) or Alaskan Coastal Current (ACC), and also to the Aleutian North Slope Current (ANSC; Stabeno et al 1999; Hu and Wang 2010). The BSC splits into two coastal currents: the Anadyr Current (AC), which flows along the west coast into the Chukchi Sea through the western side of Bering Strait and along the southwestern coastal current, which forms the East Kamchatka Current (EKC). Numerous sites of mesoscale eddy generation possibly exist within the AS and BSC, due to the interaction between baroclinic instability and the continental slope (Wang and Ikeda, 1997; Mizobata et al., 2002, 2006, 2008). The basic schematic surface circulation pattern is fairly well known (Fig. 1), based on available observational evidence (Kinder and Coachman, 1978; Stabeno and Reed 1994). The interaction or exchange between shelves and deep basins is a typical phenomenon that significantly influences primary and secondary productivities (Mizobata et al. 2002; Mizobata and Saitoh, 2004).

The Western Arctic Ocean, with a freshwater and heat pathway to the Bering Sea via Bering Strait, includes the Eastern Siberian Sea, Chukchi Sea, and Beaufort Sea (Fig. 1).
Significant change in sea ice cover and ice-related physical and ecological phenomena occurred in the last decade, particularly in the Pacific Arctic region (PAR). In particular, the extent of Arctic sea ice reached its all time low in September 2007, shattering all previous lows (Serreze et al. 2007; Comiso et al. 2008; Gascard et al. 2008; Wang et al. 2009) since satellite record-keeping began nearly 30 years ago (Fig. 2). The Arctic sea ice extent in September 2007 stood at $4.3 \cdot 10^6$ km$^2$. Compared to the long-term minimum average from 1979 to 2000, the new minimum extent was lower by about $2.56 \cdot 10^6$ km$^2$ --an area about the size of Alaska and Texas combined, or 10 United Kingdoms. The cause of this significant ice loss was thought to be the combined effects of Arctic Oscillation (AO)-induced warming (Wang and Ikeda 2000) and exporting of multiyear ice (Rigor and Wallace 2004; Steele et al. 2004), warming trend due to greenhouse gases, and the culmination of an ice/ocean-albedo positive feedback (Ikeda et al. 2003; Wang et al. 2005b).

As the Arctic environment changes at a faster rate than the rest of the world, an emerging concern is how soon the Arctic Ocean will become ice-free in summer (Wang and Overland 2009). The diminishing summer ice cover in the western Arctic can have significant impacts on the Arctic and subarctic marine ecosystems, including a lengthened algae bloom period due to increasing absorption of solar radiation (Grebmeier et al. 2006). In the future, the present Bering ecosystems may migrate to the Chukchi Sea. The Bering mammals and salmon would migrate to Chukchi due to more open water and an abundance of biomass. Seasonal landfast ice along the Alaskan Arctic coast (Eicken et al. 2006; Mahoney et al. 2007) would melt early and form late, causing a shortening of whaling and fishing seasons for both the local community and polar bears. These emerging consequences will alter not only the ecosystems, but also the lifestyle of
the native community and commercial activity in the Arctic. Such impacts are far-reaching and inherently interdisciplinary.

Since 1995, the annual sea ice variability shows that summer Arctic sea ice continued to set one record low after another (Fig. 2). During the same time, the AO index became mostly neutral or even negative (Overland et al. 2008; Maslanik et al. 2007), suggesting a weak link between the AO and the rapid sea ice retreat in the recent years. Whenever the Arctic sea ice has reached a new minimum, searching for mechanisms responsible for the individual year’s event was appealing (Serreze et al. 2007; Nghiem et al. 2007; Gascard et al. 2008). However, there has been no convincing physical explanation that accounts for the complete series of such events (1995, 1999, 2002, 2005, 2007, and 2008). Wang et al. (2009a) proposed that the Arctic Dipole Anomaly (DA) pattern is the major forcing in advecting sea ice out of Arctic Ocean under the thinning ice conditions (which is due to long-term cumulative thermodynamic effect, or ice/ocean albedo feedback), causing a series of Arctic ice minima since 1995.

In the Bering Sea and the southern Chukchi Sea, seasonal sea-ice cover is an important predictor of regional climate (Niebauer, 1980; Wang and Ikeda, 2001). Sea-ice extent also determines ocean circulation patterns, thermal structure, water stratification, and deep convection (Wang et al. 2010b) because: 1) wind stress drag is different in magnitude over water surface than over ice surface (Pease et al., 1983); 2) the albedo of ice differs from that of water; thus, prediction of the sea-ice extent (i.e., edge) is crucial to predicting the ocean mixed layer and ocean circulation, and thus, further to predicting primary and secondary productivities (Springer et al., 1996); and 3) spring ice-melt freshwater increases stratification in the upper layer, which may enhance phytoplankton blooms. In addition, climate change, through its effect on the timing of ice melting, would determine the timing of phytoplankton and zooplankton blooms. As a
result, sea-ice conditions and the ecosystem in the Bering Sea are driven mainly by atmospheric and oceanic forcing, from tidal, synoptic, and seasonal to interannual and decadal time scales.

This chapter provides an overview of several emerging ice-related processes in the Pacific Arctic and the Bering Sea, and is organized as follows. After the introduction, the next section describes data and methods used. Section 3 investigates the Arctic Dipole (DA) anomaly and summer sea ice minima. Section 4 shows the modelling studies of ice minima and its relationship to the dynamic (wind) and thermodynamic (including ice/ocean albedo feedback) forcings. Section 5 investigates the Bering Strait heat transport and its possible relation to the DA. Section 6 discusses the modeling simulation of landfast ice along Beaufort coast using a Coupled Ice-Ocean Model (CIOM). Section 7 investigates seasonal and interannual variability of the Bering Sea cold pool using CIOM. A positive air-ice-sea feedback loop is proposed and discussed in section 8. Section 9 summarizes the conclusions and proposes future efforts.

2. Data and Methods

The average September sea ice extent, archived at the National Snow and Ice Data Center (NSIDC), was obtained from SMMR (Scanning Multichannel Microwave Radiometer) for 1978-1987 and SSM/I (Special Sensor Microwave Imager) for the period 1987 to present based on NASA Goddard algorithm (Comiso et al. 2008). The monthly NCEP (National Centers for Environmental Prediction) Reanalysis dataset from 1948 to 2010 were used to derive the EOF (Empirical Orthogonal Function) modes for individual seasons: winter (DJF), spring (MAM), summer (JJA), and autumn (SON). Oceanic heat flux via the Bering Strait was calculated using in situ shipboard measurements and satellite-measured SST across the Bering Strait (Mizobata et al. 2010). A Pan-arctic Ice-Ocean
Modeling and Assimilation System (PIOMAS, Zhang et al. 2008a, b) was used to simulate the sea ice and ocean circulation for the period 1978-2009 using daily NCEP forcing.

Another regional Coupled Ice-Ocean Model (CIOM) was used in the Chukchi and Beaufort seas, and in the Bering Sea to investigate the ice-ocean systems and the responses to the significantly climate change in the recent years. The detailed description of development of the CIOM should refer to Yao et al. (2000) and Wang et al. (2002, 2010b), which was applied to the pan-Arctic Ocean (Wang et al. 2004, 2005a; Wu et al. 2004), Chukchi-Beaufort seas (Wang et al. 2003, 2008), and the Bering Sea (Wang et al. 2009; Hu and Wang 2010). The ocean model used is the Princeton Ocean Model (POM) (Blumberg and Mellor, 1987), and the ice model used is a full thermodynamic and dynamics model (Hibler, 1979, 1980) that prognostically simulates sea-ice thickness, sea ice concentration (SIC), ice edge, ice velocity, and heat and salt flux through sea ice into the ocean. The model is being applied to the northern China seas (Q. Liu, personal comm.) and the Great Lakes (Wang et al. 2010a).

3. Leading climate forcing: Arctic Dipole (DA) pattern

The EOF analysis of the NCEP reanalysis dataset (1948-2010) shows the AO as the first mode, and the DA as the second mode (Wang et al. 1995, 2009a; Wu et al. 2006; Watanabe et al. 2006) (Fig. 3). The AO has one annular (circled) center covering the entire Arctic, producing zonal wind anomalies. The AO intensity is strongest in winter (63%), and decreases in magnitude from spring (61%), summer (50%), to autumn (49%). The DA (Fig. 3) differs from the AO in all seasons because the anomalous SLP has two action centers in the Arctic. In contrast to the AO-derived wind anomaly which is either cyclonic or anticyclonic during its positive or negative phase (Proshutinsky and Johnson 1997; Wu et al. 2006), the resulting wind
anomaly by the DA is meridional. During a positive phase of the DA (i.e., the SLP has a positive anomaly in the Canadian Archipelago and negative one in the Barents Sea), the anomalous meridional wind blows from the western to the eastern Arctic, accelerating the Trans-polar Drift Stream (TDS; see the red-dashed arrow) that flushes more sea ice out of the Arctic into the Barents and Greenland seas (Wu et al. 2006; Watanabe et al. 2006). During the negative phase of the DA, the opposite scenario occurs, i.e., more sea ice remains in the western Arctic (Watanabe et al. 2006).

Note that the DA accounts for more and more total variance from winter (13%) and spring (14%) to summer (16%) and autumn (16%). The variance ratio of DA to AO is 0.21 for winter, 0.23 for spring, 0.32 for summer, and 0.33 for autumn. This pattern indicates that the DA-derived anomalous wind plays a more important role in summer and autumn during the melting and freezing seasons, or during the thinning ice season, than winter with thicker ice. The second noticeable feature is that the orientation of the maximum wind anomalies are more parallel to TDS-Fram Strait during spring and summer than winter and autumn, indicating spring and, in particular, summer DA are more effective in advecting sea ice to the Nordic seas than the winter and autumn during a positive DA phase. The third feature is that both AO and DA have intraseasonal variations.

During the positive/negative AO (i.e., when Arctic SLP has a negative/positive anomaly), a cyclonic/anticyclonic wind anomaly occurs, indicating a sea ice divergence/convergence. The divergence (anomalous cyclonic circulation) of sea ice leads to anomalous ice export, while the convergence results in retention of sea ice inside the Arctic Ocean (Proshutinsky and Johnson 1997; Wu et al. 2006). Thus, the DA pattern is the more effective and important driver of the Arctic sea ice transport from the western Pacific Arctic to the northern Atlantic.
We conducted the cross composite analysis of both phases of the DA and the AO. A matrix was constructed based on the combination of the two leading drivers that account for about 65-76% of the total variance. Following Wang et al. (2009), the Arctic climate patterns can be defined by the following four climate states: 1) +AO+DA, 2) +AO-DA, 3) -AO+DA, and 4) -AO-DA. It was found that a positive DA is the key, regardless of the sign of the AO, to flushing sea ice out of Arctic due to its dominant meridional (southerly) wind anomaly (Maslanik et al. 2007), while the wind anomaly driven by a negative DA can retain sea ice in the western Arctic (Fig. 3). In other words, the DA is dynamically more important and more effective than the AO in driving sea ice out of the Arctic Basin. These four states can represent major atmospheric circulation patterns in the Arctic, which are the major drivers of sea ice and ocean circulation.

The importance of the air temperature advection by the DA is consistent with the recent findings by Overland et al. (2008). They concluded that the warming anomalies in the central Arctic during 2000-2007 are driven by meridional advection from the south.

Figure 4 shows a scatter plot between the summer DA index (x-axis) and winter-spring mean DA index (y-axis) for the period since 1995. The figure examines whether the persistency of the DA may cause a series of ice minima. Most of the ice minima years fall into the first quadrant, which indicates that, when the +DA persists from winter, spring, all the way to summer, more ice would be advected out of the Arctic Ocean, leading to ice minima in 1995, 2002, 2005, 2007, 2008, and 2009; the 2007 ice minimum is the all time low (see Fig. 2).

Although the –DA occurs in the winter-spring period when sea ice are normally retained in the Pacific Arctic, a strong +DA in summer can reverse the process, effectively advecting sea ice to the Nordic seas and leading to ice minima in 1999, 2005, and 2010 (4th quadrant). Therefore, the summer DA is the most critical dynamically since summer sea ice is more mobile than winter.
due to that fact that summer sea ice thickness is about 1 m thinner than the thickness in winter. Because the summer anomalies have weaker AO-related SLP signals, the DA-derived wind anomalies are more effective in reducing sea ice, primarily due to two thermodynamic effects, i.e., advecting warm air from land to the ocean and enhancing melt through a positive ice-albedo feedback.

4. Modeling Arctic sea ice minima using PIOMAS

On the basis of the above analyses, the +DA-derived wind forcing is the key to the ice minima. To further confirm such a key mechanism in the summers since 1995, in particular from 2007 to 2010, the PIOMAS (Zhang et al. 2008a) was used to simulate the sea ice and ocean circulation for the period 1978-2009 using daily NCEP forcing. Figure 5 shows the August SLP and wind anomalies from 2007 to 2010. A common feature is that DA was positive for all these summers. The differences are the magnitude and orientation of the DA-induced wind anomalies. Although the wind anomalies in 2007 were larger than 2008 in magnitude, the orientation for both years were more favorable for advecting sea ice out of the Arctic into Nordic seas via Fram Strait and the Barents Sea. Similarly, wind anomalies in 2007 and 2010 have similar magnitude and both larger than those in 2008 and 2009; nevertheless, the orientation of wind anomalies in 2010 is less favorable for advecting more ice out of the Arctic than 2007. The orientation of wind anomalies in 2009 was least favorable, since the primary ice advection is towards the Kara Sea.

Figure 6 shows that the simulated sea ice area compares well against the satellite-measurements. The correlation between the observed and modeled time series of ice extent is 0.91 in September, while it is 0.93 for January-September. In particular, the model reproduces excellent agreement with summer ice minima in 1995, 2002, 2005, and 2007, 2008, and 2009,
though not for 1999. The 2010 daily summer minimum ice extent \(4.8 \times 10^6 \text{ km}^2\) was just higher than 2007 \(4.2 \times 10^6 \text{ km}^2\) and 2008 \(4.6 \times 10^6 \text{ km}^2\). This indicates that both orientation and magnitude of +DA are the key for advection of summer sea ice out of the Arctic.

To understand in depth the ice-ocean system, particularly the ice-ocean albedo feedback, in response to the strongest, most persistent (from winter to summer) DA event in 2007 and to better understand the linkages between the rapid arctic sea ice retreat and the atmospheric changes during summer (July–September) 2007, it is helpful to examine how the changes are reflected in the NCEP/NCAR surface atmospheric forcing that is used to drive the model. The changes in SLP, surface winds, and SAT leading to and during summer 2007 are illustrated in Figure 7. It compares the 2007 atmospheric conditions with those averaged over 2000–2006, a period of relatively low summer sea ice extent during the past 3 decades. Here, the 2000–2006 average is referred to as the “recent average” and the difference between a 2007 value and the recent average value is referred to as an “anomaly” (2007 minus the recent average) (Zhang et al. 2008a, b).

As shown in Figure 7a, the arctic +DA-like SLP and surface wind anomalies were small, but were gradually built up in the first half of 2007. In July, +DA-like SLP was considerably higher in much of the Arctic Basin and lower over a large area in Russia than the recent average (Figure 7b). This is associated with stronger southerly winds in the northern Canada Basin and easterly winds along the East Siberia coast. In August and September, the well-defined +DA (Wang et al. 2009) was developed: the high SLP anomalous center was mostly confined to the Canada Basin and a low SLP anomalous center was located in the Barents Sea, producing stronger southerlies in the Pacific sector (Figures 7c–7d).
Similar to the SLP and wind anomalies, the arctic SAT anomaly is relatively small in the first half of 2007: SAT is slightly warmer in the central Arctic Basin and slightly cooler in the Chukchi and Beaufort seas compared to the 2000–2006 average (Figure 7e). In July, the SAT anomaly increases significantly in the Chukchi Sea (Figure 7f). In August and September the increase in SAT is most striking: SAT is up to 5°C warmer than the recent average over much of the Pacific sector (Figures 7g–7h). However, in an area in the Canada Basin, SAT is lower than the recent average in September 2007 (Figure 7h). This SAT decrease is likely due to the air–ice interactions in that area where ice is thicker than the recent average. Sea ice is also thicker in August in that area (Figure 7k); however, SAT is not lowered (Figure 7g).

What is the cause of the substantially reduced ice cover in the Pacific sector during summer 2007? Zhang et al. (2008a, b) found, based on model results, that part of the anomalous reduction (~30%) in ice extent mainly in the Pacific sector is due to the unusual ice advection, which was caused by the strengthened TDS by the DA-derived meridional wind anomalies. Ice motion in the Arctic Ocean is characterized by an anticyclonic Beaufort gyre and the TDS (Figures 8a–8d). In the first half of 2007, the SLP and wind anomalies are small (Figure 7a), so the changes in ice motion and changes in ice thickness due to ice advection are small in comparison to the 2000–2006 average (Figure 8e). In July, there were stronger southerly winds in the northern Canada Basin (Figure 7b). The ice motion responds to the winds with a much stronger TDS, so that there are considerable changes in ice thickness due to ice advection almost everywhere (Figure 8b). In a large area of the Pacific sector, the reduction in ice thickness due to ice advection is up to 0.5 m/month more than the recent average (Figure 8f). In August and September, there were even stronger southerly winds in the Pacific sector (Figures 7c–7d), which further strengthen the ice motion and TDS (Figures 8g–8h). Thus, the Pacific sector continues to lose ice. Again, the
reduction in ice thickness due to ice advection is up to 0.5 m/month more than usual (Figures 8g–8h). In other words, from July to September 2007, the unusual ice motion pattern drives so much ice into the Atlantic sector from the Pacific sector that the ice thickness in most of the Pacific sector is reduced by up to 1.5 m more than the recent average, leaving the Pacific sector with an unusually large area of thin ice and open water.

Major anomalous reduction (~70%) in ice extent in the Pacific sector during summer 2007 is due to reduced ice production or enhanced ice melting (Zhang et al., 2008a), which includes the ice-ocean albedo feedback (Wang et al. 2005b) and ice-cloud feedback (Ikeda et al. 2003). Ice production is the net ice thermodynamic growth or decay due to surface atmospheric cooling/heating and ocean heat flux. In summer, a decrease in ice production is equivalent to an increase in ice melting. Ice production is negative in July and August due to ice melting caused by atmospheric or oceanic heating (Figures 9b and 9c); it is generally positive in the January–June mean (Figure 9a) and in September (Figure 9d). In the first half of 2007, the ice production anomaly is generally small (Figure 9e). In summer 2007, however, ice production decreases considerably in most of the Pacific sector (Figures 9f–9h) where a large reduction in ice thickness due to ice advection also occurs (Figures 8b–8d and 8f–8h). This is because that the large reduction in ice thickness due to ice advection is associated with a large area of thin ice and open water, and thin ice and open water tend to allow more surface solar heating owing to the ice-albedo feedback (Wang et al. 2005b), leading to reduced ice production or enhanced ice melting (Zhang et al. 2008a; Lindsay et al., 2009). Therefore, the positive ice/ocean albedo feedback plays a key role in the thermodynamic melting during the 2007 summer.

The changes in ice production represent the net effects of all the air–ice–ocean thermodynamic processes, which are also indirectly contributed by the oceanic and atmospheric
dynamic forcings that enhance the thermodynamic melting, such as enhancing the northward oceanic heat transport via Bering Strait (Mizobata et al. 2010; Woodgate et al. 2010) and advecting warm air temperature from the south (Overland et al. 2008). Although the reduced ice production or enhanced ice melting in the Pacific sector during summer 2007 is dominated by intensified surface solar heating because of the ice-albedo feedback (Zhang et al., 2008a,b; Wang et al. 2005b), other thermodynamic processes also play a role via both direct and indirect contribution. For example, intensified solar heating also warms the surface waters in the Pacific sector considerably (Steele et al. 2008; Perovich et al. 2008). The unusually warm surface waters in turn warm the overlying atmosphere, elevating summer SAT in the Pacific sector (Figures 7f–7h). The stronger southerly winds associated with +DA (Figures 7b–7c) may also contribute to increasing SAT by bringing warm air from the south to the Arctic (Overland et al. 2008). An increase in SAT leads to an increase in surface longwave radiation and turbulent heat fluxes, resulting in additional ice melting (Zhang et al. 2008a; Ikeda et al. 2003). Ocean circulation (advection) and its thermodynamic processes (heat transport via Bering Strait) have also played a role in the reduced ice production or enhanced ice melting in the Pacific sector during summer 2007. Therefore, strictly speaking, the 70% of ice reduction due to thermodynamic processes should also include the indirect contribution by the oceanic and atmospheric dynamic forcings.

5. Bering Strait Heat Flux and DA

The +DA not only drives sea ice from the Western/Pacific Arctic to the Eastern/Atlantic Arctic via accelerating the TDS, but also sucks the inflow of the warm Pacific water (Woodgate and Aagaard. 2005) that injected above-average heat flux from the Pacific, accelerating the drastic thinning of sea ice (Steele et al. 2004; Shimada et al. 2006; Zhang et al. 2008a). To
confirm that the Pacific water heat flux increased in the 2000s, in particular in summer 2007, we calculated the heat flux through the eastern Bering Strait from 2000 to 2010 during the June-October ice free season. Figure 10 shows that since 2004, heat flux via the Bering Strait has an annual average of 12.14 TW (1 TW=10^{12} \text{ Watts} \text{ and } 1 \text{ Watt}= 1 \text{ Joule/s}) from 2004-2007, compared to the annual average of 6.4 TW during 2000-2003, representing a 90% increase. The heat flux in 2007 (12.36 TW) had a 30.97% increase compared to the average of 9.27 TW from 2000 to 2007. The heat flux from the Pacific Ocean has two important impacts, both direct and indirect, on sea ice in the western Arctic. The direct impact includes the bottom and lateral melting of sea ice when the warm Pacific enters the Chukchi Sea, which enhances the melting via instant (seasonal) ice/ocean albedo feedback (see the discussed in section 4). The extreme event was the 2007 summer. The indirect impact involves a time-lag effect: the oceanic heat flux entering in the previous summer may survive winter (Shimada et al. 2006; Hu and Wang 2010) at the subsurface, which enhances the melting in the following spring and summer, amplifying the ice/ocean albedo process.

There is a possible relationship between the heat transport and the DA. Mizobata et al. (2010) showed that the regression model using NNW wind component well reproduced the Bering Strait inflow. Figure 11 shows the scatter plot of heat transport against summer +DA indices from 2000 to 2009, since the heat transport data are available after 2000, while the data for 2010 heat flux was not completed (Fig. 10). Within the positive territory of summer DA index, the heat flux ramped up with positive DA index, with 2007 being the record high (Woodgate et al. 2010). The reason is that the orientation of the summer DA-induced anomalous wind maximum is almost parallel to the TDS and the Bering Strait, effectively driving Bering warm water (i.e., anomalous heat flux) northward. In addition, a northward wind anomaly in the
Bering Strait could enhance the heat transport to the Arctic Ocean in August 2007 (Fig. 10b). By contrast, the heat flux in August 2008 was only 3.52 TW, much smaller than the mean (9.27 TW) of 2000-2007 (Fig. 10c). The reduced flux in 2008 is likely due to a smaller magnitude of the +DA than that of 2007 (Fig. 5b), Another important reason is that over the Bering Strait, a southward local wind anomaly (Fig. 5b) can significantly reduce the northward heat transport in 2008. Therefore, the local wind anomalies should be included as a factor of influencing heat flux (Mizobata et al. 2010). The correlation between the estimated heat flux and summer DA index is 0.51, although it is not significant at the 95% significance level (0.71) due to small number of the samples.

Although there is an association between the heat flux and +DA events, it needs to be further investigated using long-term measurements. Furthermore, in addition to large-scale and local atmospheric forcing, the orientation, bathymetry, and ocean boundary around Bering Strait may be factors affecting the northward heat transport, which differs from the atmosphere without lateral boundary.

6. Modeling landfast ice in the Beaufort-Chukchi Seas using CIOM

Landfast ice along the Chukchi and Beaufort coast is a seasonal phenomenon (Eicken et al. 2006; Mahoney et al. 2007). It is a great challenge for any coupled ice-ocean model to capture the dynamic and thermodynamic features of landfast ice, since many factors can affect the formation, anchoring, and melting of landfast ice, such as wind forcing, ocean currents, coastal topography and bathymetry, and model resolution. To this end, a 3.8-km resolution CIOM (Wang et al. 2003, 2008; Jin et al. 2007) was used to investigate the seasonal and interannual variability of landfast ice in the Chukchi and Beaufort seas.
Figure 12 shows the CIOM-simulated climatology of June sea ice concentration (SIC) and thickness using the daily NCEP forcing for the period 1990-2009. The SIC map (Fig. 12a) clearly shows what corresponds to simulated landfast ice attaching to the Alaskan Beaufort and Chukchi coast during the melt season as ice of high concentration. During spring, surface melting commences nearshore but ice concentrations first drops offshore as a result of complete melting and removal of thinner offshore ice in areas of higher open water concentrations that promote absorption of solar heat. This is also reflected in the small magnitude of ice velocity vectors superimposed on ice concentration in Fig. 12a; nevertheless, since landfast ice is not modelled explicitly, small residual velocities remain in some areas of effective landfast ice. However, at the same time, the strong contrast between stationary landfast ice and highly mobile ice just offshore from the landfast ice edge appears to be well captured. Even in mid-July, Beaufort landfast ice remains, not melting completely until the end of the month, depending on weather (SAT and wind direction) conditions.

The simulated sea ice thickness map (Fig. 12b) in June shows some contrast in landfast ice thickness along the Beaufort and Chukchi coast (~1.5m) and thinner offshore (<1m) ice. Since the model is not explicitly simulating processes that contribute to landfast ice stabilization, in particular grounding of pressure ridges (Mahoney et al., 2007), other processes represented in the model drive the mechanics of formation and maintenance of landfast ice. These include the following factors (Wang et al. 2010b): 1) a northeast wind due to the Beaufort high pressure system, 2) the eastward ACW current with its right-turning force due to the Coriolis effect, 3) high resolution topography and bathymetry constraining ice motion in the coastal regions, and 4) sea ice advection. However, at this point it is unclear how these model-inherent factors relate to the processes that help keep landfast ice in place in nature.
Figure 13 shows the climatology (1990-2009) of the simulated landfast ice that was compared to observed landfast ice extents obtained from synthetic aperture radar satellite data for the period 1995-2005 (Eicken et al. 2006; Mahoney et al. 2007). In the model, landfast ice starts to form in autumn due to the Beaufort Gyre and anticyclonic winds induced by the Beaufort High, both of which push sea ice toward the Alaskan Beaufort coast, coupled with the thermal growth of sea ice along the shore (Wang et al. 2010b). When sea ice completely covers the entire Arctic from December on, landfast ice is attached to shore, while pack ice offshore still moves with the ocean surface current and wind forcing. During the period of complete ice cover, the radar satellite data indicate completely stationary landfast ice with a clearly delineated boundary between pack ice and landfast ice (anchored to bottom and attached to shore with the velocity almost being zero), while the CIOM-simulated landfast ice still exhibits small movement since the sea ice produced in CIOM is not resolving the anchoring of grounded pressure ridge keels that stabilize the landfast ice. Thus, more research is required to improve ice dynamics representation in coastal regions and landfast ice processes by formulating and including the relevant ice anchoring mechanisms in the model.

Nevertheless, the CIOM-simulated landfast ice is mostly consistent with landfast ice extent derived from satellite data. The CIOM reproduces the landfast ice boundary in January and February very well. However the model reproduces less landfast ice than the measured boundary in March. During April, a melting season, CIOM reasonably well reproduces landfast ice in general, but reproduces less ice from 147-152W. In May, CIOM reproduces more landfast ice between 140-147W. In June, the model simulation compares very well with the measurement.
There are two approaches to distinguish landfast ice from pack ice in CIOM. One way is to define landfast ice by an ice velocity criterion that considers ice stationary below a given velocity threshold. Here, if the absolute ice velocity is at or less than 4 cm/s at water depths less than 35m, then grid cells are designated as landfast ice. The second, prescriptive method stipulates that during the simulation, the wind stress, and ice velocity are set to zero shoreward of the 35m isobaths, roughly corresponding to the extent of landfast in many areas (Fig. 12). This method is widely used in Baltic Sea ice simulations (Haapala et al. 2001; Meier, 2002a, b), but is not capable of representing spatial and interannual or seasonal landfast ice extent.

7. Modeling the Bering Sea Cold Pool using CIOM

An important seasonal feature on the Bering Shelf is the cold water (cold pool) mass on the bottom of the middle shelf (50-100 m isobaths) that persists throughout the summer (Takenouti and Ohtani, 1974; Kinder and Schumacher, 1981; Wyllie-Echeverria 1995). The cold water was even observed in late September and early October (Kinder and Schumacher, 1981). The largest vertical temperature difference, surface minus bottom, is >7 ºC in the middle domain, suggesting that the cool pool survives the summer (Hu and Wang 2010). The cold pool extends from the Gulf of Anadyr in the west with a temperature of < -1.0 ºC (Hufford and Husby 1972) to a variable eastern boundary over the southeastern shelf with temperature of 2 ºC (Maeda et al., 1967). The extent and volume of cold pool vary upon the past winter’s meteorological and oceanic (convection) conditions. The minimum annual extent can reach eastward about 170 ºW, while the maximum extent can cover Bristol Bay (Fig. 14a).

The existence of the cold pool in summer is important not only to the physical environment, but also to marine ecosystems. The cold pool is an ideal habitat to some arctic cold
water species such as arctic cod, and acts as a barrier to certain species such as walleye pollock since they prefer water temperature warmer than 2 °C. When the volume and extent of the cold pool change from year to year, the population of local species may increase or decrease accordingly to their preference. Thus, the cold pool can affect biomass growth rate and distributions on the Bering Sea shelf.

The summer (August-averaged) cold pool is reproduced by the CIOM (Fig. 14b). The cold water lies on the middle shelf between the 50-100 m isobaths, and it may reach the 200-m isobath in the western shelf break. The bottom temperature increases gradually from the 50-m isobath to the Alaskan coast. The maximum temperature reaches 11 °C near Norton Sound coast. The basin water at 200 m is basically > 3 °C. The minimum bottom temperature in the middle of southeastern shelf is > 0 °C, compared to the northwestern shelf bottom water of < 0 °C.

A vertically stable temperature structure is a major factor that insulates heat transfer from surface layer to bottom layer (Hu and Wang 2010). Figure 15 shows monthly-averaged progression of vertical temperature profiles in the cold pool region (see location in Fig. 14b). The winter temperature structure is vertically homogeneous (November to March), indicating the surface-to-bottom convection and production of winter shelf water. As solar radiation increases in March, the surface temperature rapidly increases. The upper mixed-layer gradually reduces to a minimum in July during which SST is relatively high. SST gradually increases from August to September, reaching the maximum in September. Then, the mixed-layer starts to deepen due to cooling in September. From October to November, the vertical stratification rapidly disappears during strong wind mixing and cooling, reaching vertical homogeneous in December.

It is noted that the bottom cold water, which forms during winter seasons, can survive the summer months (Fig. 15). At this specific northern location, the bottom water is as cold as -1 °C.
at the bottom, with a well-defined stratification in August and September. Therefore, the cold pool water mass is a seasonal phenomenon on the Bering shelf, which has significant impacts on regional ecosystems and fisheries distribution.

The CIOM was integrated from 1990 to 2008 using daily atmospheric forcing. The same formula of Hu and Wang (2010)

\[
Volume_{\text{CP}}(t) = \sum_{i,j,k=1}^{N \times M \times K} \Delta x_{i,j,k} \Delta y_{i,j,k} \Delta z_{i,j,k}, \quad \text{if } T(x,y,z,t) \leq 2^0 \text{C}
\]

was used to calculate the volume of the cold pool.

Figure 16 shows the interannual variability of the cold pool volume, which is an indicator of year-to-year water property change induced by atmospheric forcing. The colder the atmospheric temperature and the stronger the northerly winds, the stronger the vertical convection, leading to a larger cold pool extent and volume. The cold pool has its high extent/volume in 1991, 1992, 1995, 2000, 2002, 2006-2008, and has its low extent/volume in 1993, 1996, 2001, and 2005. The year-to-year changes in volume of the cold pool significantly impact or modify the distribution of marine ecosystems since temperature-sensitive species are distributed according to their preference of habitat environment.

It is noted that there are two processes involving the year-to-year change in the cold pool volume. The first one is the abrupt change of atmospheric forcing. For instance, abrupt decrease in the volume in 1992-93, 1995-96, and 2000-01 reflects the abrupt change in atmospheric conditions with warming and weak northerly winds. The abrupt increase in volume occurred in 1996-97, 2001-02, and 2005-06, indicating the sudden cooling and increase in northerly wind anomalies over the Bering Sea. The second process is the water temperature memory effect. During the increase progression in cold pool volume, it often takes three years to reach the maximum such as in 1990-92, 1993-95, and 1998-2000, indicating the memory of water heat
storage. However, very often time, it only takes one year to see the abrupt increase in cold pool volume such as from 1996-1997, 2000-2001, and 2005-2006, indicating a significant change of atmospheric forcing. The longest decrease in cold pool volume occurred during 2002-2005, followed by the three straight high years from 2006 to 2008, indicating a cold phase in the region, consistent with the cooling of the water temperature in the M2 site in the southeastern Bering shelf (see Fig. 5 of Overland et al. this volume). The persistency of the large cold pool volume during 2006-2008 should have significant impacts on ecosystems and fisheries in the coming years, which should be closely monitored.

8. Possible Positive Air-Ice-Sea Feedback in the Western Arctic

In the Pacific Arctic, from the Bering Sea to the Arctic Ocean, there is a regional feedback loop among the atmosphere, sea ice and ocean (Wang et al. 2005b). The recent increase in Bering Sea heat transport into the Chukchi Sea since 2000 (Woodgate et al. 2010; Mizobata et al. 2010) is closely related to the continuous ice decline in the Western Pacific (Shimada et al. 2006). However, the cause is not well understood, although a well-known positive ice/ocean albedo feedback is in play (Wang et al. 2005b).

Figure 17 depicts a possible regional interaction among the atmosphere, sea ice, and the ocean. The DA-induced wind anomaly enhances the northward transport of the Bering warm water, i.e., heat flux (Mizobata et al. 2010), which also drives sea ice from the Western Arctic to the Eastern (Wang et al. 2009) via strengthening the TDS (Comiso et al. 2008). The oceanic heat flux entering the Chukchi Sea instantly enhances the melting of sea ice, and increases the local heat storage, as discussed in section 4 and by Zhang et al. (2008a, b). Thus, the excessive heat would accelerate the melting process during the following winter and spring, and slow down the
freezing in autumn. In summer, the open water absorbs more solar radiation and the heat storage increases, which is the well known positive ice/ocean albedo feedback (since the ice and ocean albedo is \(\sim 0.7\) and 0.1, respectively). The heat stored in the upper ocean layer will be present in the following autumn and winter, increasing the SST and SAT, delaying the freezing in the autumn and thinning ice in the winter. The thin ice in winter will lead to early break up in the following spring, and to more open water in the following summer, absorbing and storing more heat from solar radiation.

In summary, +DA-induced northward wind anomaly has two important impacts on the Western Arctic ice-ocean-ecosystems. The first is a direct, seasonal impact. The northward wind not only drives sea ice away from the Western Arctic to the Eastern Arctic, but also enhances the inflow of the warm Bering water into the Chukchi Sea, instantly melting sea ice in the Chukchi Sea and increasing the SST and SAT (Zhang et al. 2008a,b; and see discussion in section 4). The seasonal ice/ocean albedo feedback also accelerates the melting in the Chukchi Sea. The second impact is an indirect, year-to-year cumulative impact. Since the heat flux advected into the Chukchi Sea increases, the excessive heat can be stored in the water column and often can survive the winter. This would lead to the positive ice/ocean albedo feedback in the following year, as depicted in Fig, 17 and discussed above.

As suggested by Ikeda et al. (2003), cloud data collected over the Arctic Basin have proved that a cloud increase during the period 1960-90 made a significant contribution to a reduction in net upward longwave radiation. The importance of clouds implies also to the decadal ice cover variability. The heat flux trend was estimated to be a linear increase in the annual flux to the ice-ocean system at 10 Wm\(^{-2}\) in the 30-year period. It is a logical expectation that part of the increase in autumn-to-winter cloud cover is in turn due to increased evaporation
from the increased open water area. Thus, an ice-cloud feedback may enhance the ice cover reduction (Fig. 18). The cloud cover changes over the Arctic induce thermodynamic effects comparable with the albedo reduction caused by the ice reduction. Since Arctic clouds are not convincingly simulated in doubled carbon dioxide experiments, further work on cloud parameterization, and subsequent analysis of model output, are highly recommended.

9. Conclusions and Future Outlook

Based on the above investigations of several emerging climate-related ice-ocean processes, the following conclusions can be drawn:

1) Although the ratio of AO to DA is about 4:1 in terms of total variance, the DA-induced wind anomaly is meridional, compared to the zonal wind anomalies induced by the AO. Furthermore, the +DA-induced maximum wind anomaly is nearly parallel to the TDS, thus accelerating the TDS and effectively enhancing ice export. There is intraseasonal variation of the DA: 1) the summer and autumn DA intensity is larger than the winter and spring, and 2) the orientation of the DA-induced wind anomaly in spring and summer is more parallel to the TDS than the winter and autumn. Thus, the summer +DA is most effective in advecting sea ice out of the Arctic than the other seasons. The persistency of the +DA from winter, spring, to summer can be used to gauge the ice minimum occurrence under the present thinning ice conditions.

2) The PIOMAS successfully captures almost all of the ice minima since 1995, which compares very well to the satellite measurements. The key factors include the synoptic daily atmospheric forcing derived from the NCEP reanalysis, where the wind forcing is the most important in recent years (under the thin ice conditions)
since summer sea ice more mobile (or sensitive to wind forcing) than ever. The modeled and measured September sea ice extents have a correlation coefficient as high as 0.91, indicating the PIOMAS captures the correct thermodynamics and dynamics of the ice-ocean systems.

3) The +DA not only drives away sea ice from the Pacific Arctic to the Atlantic Arctic, leading to ice minima, but also sucks in the warm Bering water with anomalous heat flux. In particular, the 2007 summer DA has the most parallel orientation to the TDS and Bering Strait than the summers of 2008, 2009, and 2010, thus leading to high heat transport into the Chukchi Sea. The local wind stress over Bering Strait is also important.

4) The basic patterns of landfast ice distribution, formation, and melt were captured by the 3.8-km CIOM. Satellite radar data were used to validate and evaluate the simulations. Landfast ice dynamics and thermodynamics are complicated and not well represented in many ice-ocean models. Nevertheless, the CIOM is able to simulate key aspects of the major dynamic and thermodynamic features of the landfast ice through parameterization or simplification of key processes. However, in particular the anchoring of the landfast ice by grounded pressure ridges is currently not simulated and needs to be represented for a full evaluation of coastal ice dynamics.

5) The CIOM has the capability to simulate the Bering shelf cold pool, which has significant impacts on ecosystems and fisheries distribution. The winter cold water is preserved in the middle shelf throughout the summer due to vertical insulation from the stable stratified temperature structure and also due to the lack of effective
horizontal heat transport from the deep basin and the inner shelf. Interannual variability of the cold pool extent/volume depends on winter meteorological and oceanic conditions and ocean convection.

6) The Pacific Arctic atmosphere-ice-ocean interactions can be summarized by a simple feedback loop to explain the continuous ice decline and warmer water temperature in the Chukchi Sea in the last decade. The DA plays an important role as an external forcing to the ice-ocean systems in PAR. However, once the +DA applies its forcing in the Pacific Arctic, a long-lasting impact on the sea ice reduction was observed. The key is the positive ice-ocean albedo feedback (Wang et al. 2005; Shimada et al. 2006; Zhang et al. 2008a, b). Therefore, a series of +DA should apply a series of positive pulses of positive ice-ocean albedo feedbacks to the ice-ocean systems. A possible reversal may be accomplished by −DA events. The ice-cloud feedback is another mechanism for continuous, further reduction of sea ice in the PAR.

The Arctic Ocean and the Pacific Arctic region in particular have experienced rapid changes not only in sea ice and ocean, but also in ecosystems, in particular in the last decade. This study has discussed several important processes in the region. There are urgent needs for a better understanding of these processes or hypotheses using combined long-term measurements and ice-ocean models. The following processes may be of interest:

1) There was a significant change in climate between the early 1990s and last decade. The +AO dominated in the early 1990s (cyclonic anomalous circulation), while the +DA prevailed in the last decade (meridional anomalous circulation), leading to significant ice reduction. These two leading climate modes should be the major forcing to other
subsystems, such as sea ice, ocean, and ecosystem. Therefore, to adequately address the Arctic processes, the combined impact of these two modes is necessary.

2) Bering Strait is a pathway of freshwater and heat to the Arctic Ocean and it was important even in the last decades. The connection between the Bering Strait and Fram Strait-Barents Sea opening, along with the Canadian Arctic Archipelagos (CAA) is not well known. A high-resolution coupled ice-ocean model that can resolve CAA would be a high priority.

3) With diminishing summer ice in the Pacific Arctic, a northward migration of ecosystems would be an urgent topic to be studied. The changes of food web, lower trophic level ecosystems, and fisheries can have large impacts on the future planning and management of marine resources.

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Figure Captions:

Figure 1. Topography and bathymetry of the Bering Sea and the western/ Pacific Arctic region (PAR), and schematic circulation systems (by colored arrows). Water depths are in meters. (courtesy of Weingartner, University of Alaska Fairbanks).

Figure 2. Time series of average September sea ice extent from satellite measurements from 1979 to 2009 and the spatial distribution of the recording-breaking minimum sea ice extent on September 14, 2007 (all time low).

Figure 3. Intraseasonal variations of regression maps of the first two leading modes (AO and DA) to the winter, spring, summer, and autumn mean Northern Hemisphere SLP field using the NCEP Reanalysis dataset from 1948 to 2010. Contour intervals are 0.5 hPa (see color bars). The black arrows on the left column indicate the cyclonic (anticlockwise, divergent) wind anomaly during the +AO phase (which promotes advection of sea ice out of Arctic via Fram Strait) and anticyclonic (clockwise, convergent) wind anomaly during the –AO phase. On right column, the black arrows indicate that the wind anomaly blows from the western to the eastern Arctic during the +DA phase that accelerates the TDS (in red-dashed arrows), and vice versa during the –DA phase that slows down the TDS.

Figure 4. Scatter plot between the summer DA index (x-axis) and winter-spring mean DA index (y-axis) from 1995 to 2010. The box outlined by dotted lines indicates the 0.6 index threshold; the indices greater than 0.6 are considered significant DA years.

Figure 5. The 2007 (a), 2008 (b), 2009 (c), and 2010 (d) August SLP (shaded) and wind (vectors) anomalies relative to the 1948-2008 mean (data from NCEP Reanalysis). Red/blue indicates the positive/negative anomalies in SLP. The green arrows indicate the +DA-derived anomalous wind velocity (in ms$^{-1}$). The DA-derived wind anomaly was the dominating driver for advecting sea ice toward the eastern Arctic, leading to the record minimum in the Arctic.

Figure 6. Comparison between the PIOMAS-simulated and SSM/I-observed ice extent for the period 1978-2009.

Figure 7. Anomalies of the NCEP/NCAR reanalysis SLP and surface wind (a–d) and SAT (e–h); anomalies of simulated sea ice thickness (Hi) (i–l) In Figs. 7-9, an anomaly is defined as the difference between the 2007 value and the 2000–2006 average. One of every 36 wind vectors is plotted in (a–d). The green line in (e–h) represents satellite observed ice edge and yellow line in (i–l) model simulated ice edge.

Figure 8. PIOMAS-simulated ice motion (vectors) and ice advection (IA) (color contours) (a–d) and their anomalies (e–h). Note the difference between ice motion and ice advection. Ice motion is described by ice velocity and ice advection by ice mass convergence [$-\nabla \cdot (\mathbf{u} \mathbf{h})$], where $\mathbf{u}$ is ice velocity and $h$ is ice thickness. One of every 36 ice velocity vectors is plotted.
Figure 9. PIOMAS-simulated ice production (IP) (a–d) and anomaly (e–h). Ice production is the net ice thermodynamic growth or decay due to surface atmospheric cooling/heating and oceanic heat flux, including ice-ocean albedo feedback contributed by the warm Bering water.

Figure 10. a) Volume transport through the eastern Bering Strait for the period 1999-2010. Dotted lines indicate mean volume transport between 1999 and 2003 and between 2004 and 2007. Two-way bars close to x-axis show period from June to October. Numbers above two-way bars show the integrating total volume transports from June to October through the entire eastern channel; (b) Time series of averaged sea surface temperature (gray) and NCEP2-derived 330° wind component (Wnd330ip) of sea surface wind (black); and (d) estimated heat flux through the eastern channel of Bering Strait from 1999 to 2010. Black bar means heat flux through Area 1, while white bar indicates heat flux through Area 2, which make up the entire eastern Bering Strait channel. (see Mizobata et al. 2010 for detail).

Figure 11. Scatter plot between the summer DA index (x-axis) and northward heat transport via the eastern Bering Strait (y-axis) from 2000 to 2009. Units is in 10^19 joules/year (June-October). The correlation between the heat flux and summer DA index is 0.51, which is lower than 0.71, the 95% significance level.

Figure 12. The CIOM-simulated June climatological sea ice concentration (a, from 0 to 1) with sea ice velocity vectors superimposed (vector size indicates speed in m/s), and sea ice thickness (b, in meters). Landfast ice remains attached along the Beaufort Sea coast. The green (orange) arrows denote the ocean surface current (ice flow).

Figure 13. The CIOM-simulated January-June climatological landfast ice extent (from 1990-2009, black shaded) compared to landfast ice edge locations derived from synthetic aperture radar satellite data (red dots) averaged for the period 1996-2004. Green vectors: Wind stress in 10^-5 m^2/s^2.

Figure 14. The measured cold pool extent (a) and the CIOM-simulated summer bottom water temperature (b). The star indicates the location of (173.5W, 62.7N) used in Fig. 15.

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Figure 17. A proposed possible positive air-ice-sea feedback in the western Pacific Arctic Region. +/- indicates positive/negative feedback.

Figure 18. Negative feedback between ice cover anomaly and clouds. Less ice cover (or more open water) leads to more clouds cover in fall to spring, which further results in less ice cover due to reduction of long-wave radiation to the upper atmosphere (or space). The long-wave radiation is reflected back to the ice-ocean surface, leading to higher SAT/SST anomalies.
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