Responses of surface heat flux, sea ice and ocean dynamics in the Chukchi-Beaufort Sea to storm passages during winter 2006/2007: A numerical study

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Abstract

This study investigates storm impacts on sea ice, oceanic dynamics, and surface heat flux in the Chukchi- Beaufort Seas using a coupled ice-ocean model (CIOM). Two types of storms affect the study area: Arctic-born cyclones and north-moving Aleutian lows, which lead to strong westerly and easterly winds, respectively. Driven by 6-hourly forcing, the CIOM successfully reproduced the storm impacts. Simulated sea ice movements are comparable to the satellite observations. Storms usually result in fast-moving ice, which is faster than the speed of the surface water, and the gap between sea ice and surface water speed increases with wind speed. Storms can alter the pathways of the Pacific inflow water: with westerly winds, the Pacific inflow water goes no further than the latitude of Wrangel Island, while with easterly winds the Pacific inflow water can flow northward into the interior Arctic basin. Strong easterly winds associated with north-moving Aleutian lows reverse the Alaskan Coastal Current and the Bering

Slope Current, and induce upwelling along the north Alaska coast. Westerly winds associated with Arctic-born cyclones act in an opposite way. During the storms, heat loss to the atmosphere is about twice that of normal conditions, which is mainly attributed to increases of the sensible and latent heat fluxes over the open water. Heat loss over ice was quite stable with some small fluctuations in response to the storms.

Keywords: Storm impacts; sea ice; surface heat flux; ocean circulation; upwelling and downwelling; landfast ice; coupled ice-ocean modeling; the Chukchi and Beaufort Seas

1. Introduction

The Chukchi-Beaufort Seas (Fig. 1) have several important hydrological features during winter: they provide the pathway of the Pacific inflow water to the Arctic, contain the Alaskan Coastal Current (ACC), have seasonal to perennial sea ice cover, and contain land-fast ice along the Alaskan coast. The minimum ice extent and edge in this area have been widely used as indicators of global climate change (Wang et al. 2009; Stroeve et al. 2011).

The Bering Strait/Chukchi Sea is one of the three pathways of northward-propagating cyclones into the Arctic during winter (Serreze and Barry 1988; Zhang et al. 2004). The northward moving cyclones not only transport heat and moisture to the Arctic, but also have significant impacts on the ice-ocean system underneath. Two types of storms are known to influence the Chukchi and Beaufort Seas: Arctic-born and Pacific-born storms (Pickart et al. 2013). The Arctic-born cyclones often result in westerly winds in the Chukchi-Beaufort Sea,

while the northward moving Pacific-born storms, Aleutian lows, often result in easterly winds. These two types of storms may have different impacts on the Chukchi-Beaufort Sea. As the climate warms, it is predicted that the number of high-latitude storms will increase and strengthen in intensity (Zhang et al. 2004; Sorteberg and Walsh 2008).

Storm forcing on the marginal seas of the Arctic Ocean is of importance in both ocean and sea ice circulation and air-sea-ice interactions. The shelf circulation, sea ice movements, Pacific inflow, and the thinner seasonal ice cover are easily altered by strong winds (Winsor and Chapman 2004; Spall 2007). The changes in the coastal currents and upwelling/downwelling events on the continental slope are related closely to the passage of storms (Pickart et al. 2009a; b), though the presence of ice cover may reduce these impacts during winter. Storms also can lead to fast moving ice, a divergent sea ice drift, and a breakout of land-fast ice (Hibler 1980; Brummer et al. 2008; Heil 2006).

While the ice cover can significantly reduce turbulent heat fluxes from the surface to the atmosphere, openings in the ice (i.e. leads and polynyas) can lead to large heat fluxes, especially during storm passage. In leads that persist throughout the winter, relatively warm ocean water is exposed to the cold atmosphere releasing heat and moisture into the atmosphere. Bourassa et al. (2013) stated that, in the polar region, storm-driven turbulent heat and moisture fluxes may exceed in magnitude all other terms in a regional Arctic surface energy budget, and often dominate seasonal or even annual averages. However, the magnitude and variations of surface fluxes are poorly known due to the lack of observations and uncertainties in numerical model flux parameterizations (known as bulk flux algorithms). The flux parameterizations in the temperate regions are not necessarily applicable to high-latitude conditions. Observations are needed for improving the estimation of air-sea and ice heat fluxes and their parameterizations.

Even with many uncertainties in heat fluxes parameterizations, coupled ice-ocean model can still be used to examine storm impacts on surface heat flux over various surfaces.

In this paper, we use a regional high-resolution coupled ice-ocean model (CIOM) to investigate the storm impacts on the complicated ice-ocean system in the Chukchi-Beaufort Sea during winter 2006/2007. The paper is organized as follows. The CIOM and the forcing data are briefly described in section 2. Section 3 provides an overview of the storms that occurred during winter 2006/2007. Section 4 evaluates the performance of the CIOM. Section 5 explores storm impacts on several important physical processes, such as surface heat flux, sea-ice movements, the Pacific inflow in the Chukchi shelf, the ACC, and upwelling/downwelling events. Conclusions are summarized in Section 6.

2. Model and Data

2.1 Model

The model used in this study is a coupled ice-ocean model, which is widely used in Arctic, Bering Sea, and Great Lakes ice-ocean modeling (Wang et al. 2003a,b, 2005, 2009, 2010; Long et al. 2012). The ocean model is the Princeton Ocean Model (Blumberg and Mellor 1987). The sea ice component of the model is a thermodynamic model based on multiple categories of ice thickness distribution function (Thorndike et al. 1975; Hibler 1980) and a dynamic model based on a viscous-plastic sea ice rheology (Hibler 1979). In this study, ten ice categories (0, 0.2, 0.5, 1, 1.5, 2, 3, 4, 5, and 6 m) are used, each having a percentage in a grid point.

To resolve the important coastal current and eddy activities along the shelf break of the Chukchi-Beaufort Sea, Wang et al. (2008) and Jin et al. (2008) configured the model with a high resolution with $1/28^\circ$ longitude and $1/8^\circ$ latitude, which is less than the first baroclinic Rossby radius of deformation of about 21.5 km (Shaw and Chao 2003). The model domain covers

66.5°N to 75°N and 176°E-125°W. The ocean model has 24 terrain-following vertical layers with higher resolution near the surface and bottom to better resolve the surface and bottom boundary layers. The open boundaries (velocity, temperature, and salinity) are provided by a global climate (atmosphere-ice-ocean-land) model with a resolution of $1/6^\circ \times 1/4^\circ$ (Watanabe et al. 2006). The ocean model has split modes. The 2-D external mode uses a time step of 20 s, and the 3-D internal mode uses a time step of 400 s, which is also the time step of the sea ice model.

Wang et al. (2008; 2014) and Jin et al. (2008) validated the model results and found that the model successfully reproduced the ocean and sea ice circulation and captured the seasonal variation of sea ice area as compared with remote sensing data.

In this study, we used six-hourly atmospheric data from the NCEP-DOE reanalysis 2 dataset (NCEP2; Kanamitsu et al. 2002) for the time period of 1 September 2006 to 30 April 2007 as atmospheric surface forcing for the CIOM. This forcing includes 10-m winds, 2-m air temperature, 2-m specific humidity, total clouds, and downward solar radiation. The initial conditions were taken from an extended model simulation from 1990 to 2006 using NCEP2 daily forcing. The six-hourly CIOM output from 1 December 2006 to 31 March 2007 was used to study the storm impacts.

2.2 Measurements

Satellite remotely sensed datasets were used to validate the CIOM. For sea ice concentration comparisons, we used sea ice concentration data derived from the Special Sensor Microwave Images (SSM/I) on board the Defense Meteorological Satellite Program (DMSP) F-13 (Cavalieri et al. 1996). The product is available at the National Snow and Ice Data Center (NSIDC) website (<http://nsidc.org/data/nsidc-0051>).

Daily and monthly gridded sea ice motion data are used to validate the simulated sea ice movements during winter 2006/2007. The sea ice motion vectors are derived from images of several satellite remote sensors (SMMR, SSM/I-SSMIS, AMSR-E, AVHRR), the International Arctic Buoy Program (IABP) buoy data, and NCEP/NCAR Reanalysis 1 wind data (Fowler et al. 2013). The gridded fields are projected to Northern and Southern Hemisphere EASE-Grids, which has a uniform resolution of 25 km. Sea ice motion vectors from November 1978 through December 2012 are available at the NSIDC website (http://nsidc.org/data/docs/daac/nsidc0116_icemotion.gd.html).

3. Observed storms during winter 2006/2007

During winter, the Chukchi-Beaufort Sea is usually occupied by a high pressure system, the Beaufort High (Overland 2009). To its south, the North Pacific Ocean is dominated by a progression of low pressure systems propagating from west to east (Overland and Hiester 1980). As these lows progress eastward, they often intensify further in the region of the Aleutian Island chain (Wilson and Overland 1986; Terada and Hanzawa 1984; Zhang et al. 2004) and develop into powerful storms--Aleutian lows. In some cases, Aleutian lows may move northward and even progress into the Arctic region (Pickart et al. 2009a,b), which has wide-ranging influences on the Chukchi-Beaufort Sea.

During winter 2006/2007, the Chukchi and Beaufort Seas were affected by six strong storms with domain-averaged wind speed greater than 10 m/s (Fig. 2). Besides inducing remarkable changes in winds, the storms also caused significant changes in temperature (Fig. 2). Warming or cooling depends on the origination of the system. Table 1 lists the selected six cyclone cases.

The first four are northward moving Aleutian lows. The cyclones were generated in the

northwestern Pacific and Gulf of Alaska (Fig. 3), the two primary cyclogenesis locations (Wang et al. 2004).

The first one occurred during 17-21 December 2006. This was a set of cyclones that impact the Chukchi and Beaufort Seas (Fig. 3, red, pink and orange lines). On 14 December 2006, a cyclone was generated over the western Bering Sea east of Kamchatka Peninsula. The cyclone moved northeastward, arriving at the central Bering Sea on 17 December 2006 (Fig. 3, red line). The cyclone developed into a low system covering the whole Bering Sea and Gulf of Alaska with two centers located over the central Bering Sea (976 hPa) and the Gulf of Alaska (972 hPa) (Fig. 3, red and orange lines). The Chukchi Sea and the Southern Beaufort Sea was under its front flank with warm air and strong northeasterly and easterly winds. On 18 December, the center over Gulf of Alaska moved north to the Alaska, then moved northeast and disappeared on 19 December (Fig. 3, orange line). At the mean time, a strong cyclone developed over the southern Gulf of Alaska with a center pressure of 952 hPa. Then it took a similar track to the one on 18th December (Fig. 3, pink line). On 20 December, the cyclone was over the Alaska with a center pressure of 964 hPa. From 18-21, the Chukchi Sea was on the left front of the low system with strong northerly winds; the Beaufort Sea was on the right front of the low system with warm air and strong easterly winds (Fig. 4).

The second one occurred during 10-15 January 2007 (Fig. 3, dark and light green lines). The cyclone originally developed over the Kuroshio Extension south of Japan with a center pressure of 984 hPa on January 6. During the following days (January 7-10), the cyclone intensified and moved northeastward. On 10 January, the cyclone entered the western Bering Sea with a center pressure of 968 hPa. The cyclone lingered at the same location till 18 GMT January 12 (Fig. 3, dark line). At this time, the Chukchi Sea is under the cyclone's right front flank with

warm air and strong easterly winds. Then, on January 13, the cyclone expended into an Aleutian low system covering the whole Bering Sea and part of Gulf of Alaska with three low centers: one over Bering Strait (not shown), the other over Kamchatka Peninsula; the third one over south of Aleutian Islands. The front of the Aleutian low covered the Chukchi Sea and the southern Beaufort Sea. The strong pressure gradients between the low system and the high system over the central Arctic produced strong easterly winds over the study area (Fig. 4). The low center over south of Aleutian Islands intensified and moved northward approaching Alaska during January 14-15 (Fig. 3, light green line).

The Third and fourth one occurred during January 31 to February 4, 2007 (Fig. 3, dark and light blue lines). On January 28, a cyclone formed over the south of Gulf of Alaska with a center pressure of 972 hPa. The cyclone moved northwest across the Bering Sea, arrived at the south coast of Siberia on January 31 (Fig. 3, dark blue line). Meanwhile a cold anticyclone occupied the Arctic with a center (1044 hPa) located over the southern central Arctic (Fig. 4). The pressure gradient between them produced strong easterly winds in the Chukchi and Beaufort Seas. At this stage, due to no obvious warm advection, the domain-averaged air temperature did not increase. On February 1, the cyclone moved over the Siberia land and began to weaken and dissipate, leaving a warm tongue extending from Bering Sea to the Chukchi Sea and east Siberia Sea (not shown). Within the same low-pressure system, on January 31, another cyclone formed over the south of Gulf of Alaska (Fig. 3, light blue line). This cyclone traveled north across the Bering Sea and Siberia, entered the east Siberia Sea. After landed on Siberia, the cyclone began to weaken with a center pressure of 984 hPa, which then increase to 992 hPa over the east Siberia Sea (Fig. 3, light blue line). These two cyclones opened a north-south channel transporting huge amount of warm air to the Arctic. Strong southeasterly winds prevail over the study area (Fig. 4).

The averaged SAT over the Chukchi and Beaufort Sea increased from -25 to -10°C . The surface air temperature over the southern Chukchi Sea even increased to above 0°C .

The fifth one occurred during 25-28 February 2007 (Fig. 3, purple and violet lines). Two weak cyclones merged and intensified to produce this cyclone. On Feb 23, 2007, a cyclone formed over the Sea of Okhotsk, traveled northeastward across the Siberia and entered the Chukchi and the Beaufort Seas on Feb. 25 (Fig.3, purple line); the other developed over the Great Slave Lake region in northwestern Canada and traveled northward passing over Victoria Islands, then moved westward into the northern Beaufort Sea (Fig. 3, violet line). This cyclone and the former one merged into a stronger cyclone at 0 UTC on 26 February with a center pressure of 1004-hPa. The merged cyclone intensified continually and reached its peak at 18 UTC on 26 February with a center pressure of 1000-hPa. The cyclone stayed over the northern Beaufort Sea from 26 to 28 February slightly wandering, began to move eastward and weakened on February 28, and entirely left the Beaufort Sea by February 29. The merged cyclone occupied the whole Chukchi-Beaufort Sea, and strong westerly winds in the southern flank of the cyclone prevailed over the study domain (Fig. 4). Unlike the north moving Aleutian lows, this local intensified cold-core cyclone led to a decrease in the surface air temperature in the Chukchi and Beaufort Seas (Fig. 2).

The final one occurred during March 25~30 (Fig. 3, black line). On March 21, a cyclone formed over east of Greenland with a center pressure of 972 hPa. The strong cyclone traveled north over the ocean, entered the central Arctic on March 25 with a center pressure of 964 hPa. Then the cyclone moved between there and the Chukchi/Beaufort Seas during March 26-28, and left the region on March 29. It is a very strong cyclone covering almost the whole Arctic (Fig. 4). The Chukchi/Beaufort Seas was under its south flank with strong westerly and southerly winds.

Due to its origin, this storm brought lots of warm and humid air and resulted in a rapid surface air temperature increase from -30° to -13°C (Fig. 2).

4. Performance of the CIOM model

4.1 Sea ice concentration and movements

The comparison of daily sea ice area derived from SSM/I measurements and the CIOM simulation indicates that the CIOM model basically reproduces the maximum ice area and timing of ice freezing/melting (Fig. 5a). However, the simulated sea ice suddenly melted in late June due to the imposed lateral melting parameterization. The ice-melting rate is fast during June to August when compared with the SSM/I measurements. During winter, the Chukchi-Beaufort Sea was entirely covered by sea ice with concentrations larger than 0.85 in most of the domain (Fig. 5b), except for December, in which ice concentration is about 0.5-0.7 in the southern Chukchi Sea (not shown).

The daily observed and modeled domain-averaged sea ice speed during winter 2006/2007 (Fig. 6) has a high correlation of 0.8. The modeled sea ice speeds are in good agreement with the observations, though there are still some discrepancies. For example, the modeled sea ice speeds during storms on 10-13 January and 31 Jan-2 February are larger than observed. Fig. 7 shows the NSIDC observed and the CIOM simulated sea ice movements on 19 December 2006 and 27 March 2007, both during the peak of a storm. Driven by the prevailing winds associated with the storm (Fig. 4), on 19 December 2006 sea ice primarily moves westward in the northern Chukchi Sea and Beaufort Sea, and southward in the southern Chukchi Sea. For the storm that occurred on 27 March 2007 the sea ice moves primarily eastward. These results suggest that the CIOM can successfully capture synoptic events in sea ice movement.

4.2 General ocean circulation

The CIOM successfully reproduces the major current systems, which comprise of the inflow of Pacific water from the Bering Strait and the Beaufort anti-cyclonic gyre (Fig. 8). The Pacific inflow separates into three branches steered by the topography in the Chukchi shelf, as observations have suggested (e.g. Winsor and Chapman 2004; Weingartner et al. 2005; Spall 2007), one of which is the ACC (Coachman et al. 1975). The other two branches flow northward along the Herald and Central Valleys (Fig. 8, and see Fig. 1 for bathymetry). The northward flowing Herald Valley branch reaches about 72.5°N, where the shelf break is located, and part of the current turns right when it passes the line of Herald Shoal, joining the Central Valley branch. The Central Valley branch also flows northward, reaches around 73°N, and then turns right, flowing eastward along the shelf break.

The ACC flows along the northwestern Alaska coast and then enters the Barrow canyon. Some of the Barrow Canyon outflows flow along the Beaufort shelf break and slope (Weingartner et al. 2009). The vertical section along 156.5°W within the Barrow Canyon shows the ACC flowing eastward with winter mean zonal velocity > 14 cm/s (Fig. 9a). The speed of the ACC is consistent with observations: direct current measurements from October 1986 to August 1987 recorded a mean speed of 13-16 cm/s along the axis of the canyon and a somewhat stronger flow of 17-23 cm/s on the shoreward wall of the canyon (Aagaard and Roach 1990).

To the east, along the section in the direction of the SBI (Western Arctic Shelf–Basin Interactions) mooring array deployed across the Beaufort slope at 152°W (Grebmeier and Harvey 2005, see Fig. 1 for the mooring position), the simulation shows a narrow eastward-flowing current with a core speed of approximately 6-8 cm/s is located at the shelf break near 100m depth (Fig. 9b). It is called “slope current”. The current exists along the edge of both the Chukchi and Beaufort Seas (Mountain et al. 1976; Aagaard 1984; Pickart 2004; Mathis et al.

2007; Llinas et al. 2008; Spall et al. 2008; Nikolopoulos et al. 2009). Nikolopoulos et al. (2009) suggested that at most only about 30% of the Bering Strait inflow ends up in the Beaufort slope current. The direct currents observations during winter 2002-2003 reveal that, in the mean, the slope current is a subsurface feature trapped to the shelf edge near 100m depth. It is very narrow (width scale 10–15 km) with a core speed of approximately 15 cm s^{-1} and a deeper “tail” of $2\text{--}3 \text{ cm s}^{-1}$ extending to 250 m depth (Fig. 9c, from Nikolopoulos et al. 2009). Compared to the observations, the simulated slope current is narrower and weaker, extending only to 150 m depth with no deeper tail. The simulated slope current is confined to the south of BS3 station (71.39°N , 152.05°W), while the observed one extends at least to BS4 (71.44°N , 152°W) near 100m depth (Fig. 9c). Also, the observed westward flow in the upper 50m was not properly reproduced. The vertical thermo-structure along the section is reasonably reproduced. The observations and simulation all show three different water masses: Pacific water (shallower than 100m on the inshore side of the section), winter-transformed water of Pacific origin (generally shallower than 150m offshore of the shelf break), and Atlantic water (deeper than 150m) (Nikolopoulos et al. 2009) (Fig. 9b and c).

To the north of the slope current is the westward-flowing southern limb of the wind-driven Beaufort Gyre. In the inner shelf, the near shore Beaufort Sea (the shelf between the coast and the $\sim 20 \text{ m}$ isobath) is occupied by very weak under ice westward flows with zonal velocity around $1\text{--}2 \text{ cm/s}$ (Fig. 8). The year-round direct current measurements in the near shore Alaska Beaufort Sea between 1999 and 2007 showed that the mean alongshore velocities are usually not significantly different from zero over the entire record or in individual seasons, and when significant, they are small with mean alongshore velocities being less than 3 cm/s , as the near shore ice cover inhibits the transfer of momentum from the wind to the ocean (Weingartner et al.

2009). The mean current speeds for two moorings during winter 2006-2007 were 1.2 and 1.5 cm/s, respectively.

5. Storm impacts

5.1 Sea ice and surface water movements

As expected, both ice and water speed increase significantly as expected during the storms. At the peak of most storms, the ice speed in the central Chukchi Sea is about 20-30 cm/s, and > 40 cm/s during the storms which occurred during 10-15 January and 31 January-1 February 2007, while it is only approximately 12-15 cm/s in the central Beaufort Sea, i.e., sea ice moves faster in the central Chukchi Sea than the central Beaufort Sea during storms (Fig.10 a, b, red lines). Surface water speed in the central Chukchi Sea increases to about 15-20 cm/s at the peak of storms, while in the central Beaufort Sea the surface water speed is only about 5 cm/s, with little fluctuations in response to the storms (Fig. 10 a, b, blue lines). During the storms, the ice reduction factor (ratio of ice speed to wind speed in percent) is around 2.0-2.5 in the central Chukchi Sea and around 1.0-1.5 in the central Beaufort Sea (not shown). The ice reduction factor does not change much with the wind speed during the storms. During weak wind days, the 2% rule does not apply, as ocean currents dominate the sea ice drifting.

A noteworthy phenomenon is that the sea ice moves much faster than the surface water during the storms, and the gap between them increases with the wind speed (Fig. 10c). The correlation between the wind speed and the ice and water speed difference is as high as 0.87. The reasons are as follows. Sea ice is directly driven by the wind, while the water under ice feels a much-reduced stress. According to the squared law, the stress on the ice surface (or open water) is proportional to the squared wind speed (Eq. 2 and 5 in Appendix A). For example, a 10 m/s wind will have a stress of 0.16 N/m^2 on ice or open water, which four times that with a 5 m/s

wind (0.04 N/m^2). When the 1 February storm was at its strongest, the wind, ice, and water speed averaged over the central Chukchi Sea was about 28 m/s, 50 cm/s, and 32 cm/s, respectively (Fig. 10a). The calculated stresses on ice and under-ice water are 0.98 N/m^2 and 0.18 N/m^2 , respectively. The latter is one fifth of the former, i.e., the stress on the under-ice water is significantly less than the stress on ice during the peak of the storm.

The angle between the surface wind and sea ice velocity varies with wind speed. When wind is weak ($< 5 \text{ m/s}$), the angle is scattered due to the weak wind effects on the ice movements. When the wind becomes stronger, the angle tends to around -20° to -40° (negative means sea ice moves to the right of the wind), which is consistent with previous studies (Nansen 1902; Thorndike and Colony 1982; Omsted et al. 1996; Hu et al. 2011).

5.2 The pathways of the Pacific inflow

Storms can substantially alter the general shelf ocean circulation. For a typical shelf depth of 50 m and bottom friction coefficient of $5 \times 10^{-4} \text{ m/s}$, a new steady state is expected to be fully established after roughly 2–3 days (Winsor and Chapman 2004). Storms altered the pathways of the Pacific inflow water in the Chukchi Sea shelf, the response of which was dependent on the direction of the wind.

Three cases are selected to show storm impacts: the first two occurred on 17–21 December 2006 and 10–15 January 2007, in which north-moving Aleutian lows affected the domain with easterly winds over the Chukchi-Beaufort Sea (Fig. 4). The third occurred on 26–28 March 2007, which was a cyclone that covered almost the entire Arctic Ocean with associated westerly winds over the domain.

The northeasterly winds over the Chukchi Sea during the first case are similar to the winter mean wind direction though larger in magnitude (Fig. 11a and Fig. 8). The simulated pathways

of the Pacific inflow water are also similar to the winter mean (Fig. 11a and Fig. 8) with minor differences, including an enhanced southward flow along the Siberian coast, two recirculation cyclonic gyres in the west of the Herald Canyon branch of the Pacific inflow, and an obvious westward flow in the near shore Beaufort Sea east of Pt. Barrow. The strong easterly winds over the Chukchi Sea during the second case profoundly change the pathways (Fig. 11b): the Pacific water flows northward directly into the interior basin, and the eastward flowing current along the shelf-break disappears. The flow along the western Alaska coast separates at Cape Lisburne and moves northward. A weak southwest coastal current flows between Cape Lisburne and Point Barrow, and a strong westward flow occurs in the near shore Beaufort Sea. Driven by westerly winds during the third case (Fig. 11c), except for the Herald Canyon branch, most of the Pacific inflow goes no further than the latitude of Wrangel Is., and turns eastward. The ACC is very strong in this case.

Fig. 12 shows the time series of zonal water velocity at a site within the Barrow Canyon, indicating the variability of the ACC. Due to the Pacific inflow, the current within the Barrow Canyon flows to east most of time, even though climatologically easterly winds occur in winter. However, the current may reverse occasionally under strong easterly winds, as suggested by Weingartner et al. (1998) based on the observations. For an example, a strong westward flow occurred in 11-13 January 2007. Similar conditions occurred on 7-8 December 2006, 31 January to 1 February 2007, and 15-16 February 2007. Strong westerly winds accelerate the eastward flow. During 25-29 March 2007, southwesterly winds accelerated the eastward flow. The flow has two cores of high speed; one near the surface, and another located at a depth of around 50-80 m.

Winsor and Chapman (2004) explored the sensitivity of the pathways of the Pacific inflow water to the wind direction using a two-dimensional barotropic ocean-only model. The responses in the CIOM to the storms are somewhat similar to their experiments. Specifically, the coastal current along the western Alaska coastline flows tightly along the coast when westerly winds prevailed over the Chukchi Sea, while it separates from the coast when forced by easterly winds.

5.3 Beaufort Slope Current and upwelling/downwelling events

During autumn and winter, the Beaufort slope current becomes bottom trapped, but often reverses (flows westward) under strong easterly winds (Llinas et al. 2008). As discussed in Section 4.2, the mean structure of the slope current is basically reproduced by the CIOM (Fig. 9b). Fig. 13a shows the variations of the current during the whole winter. In normal days with weak along-shore easterly winds, the current was an undercurrent with a zonal velocity of about 10 cm/s. During storm days with westerly winds, the zonal velocity of the current could be > 30 cm/s, such as the periods 25 February-2 March, and 25-29 March 2007. The current may reverse sometimes under strong easterly winds. For example, a strong westward flow occurred on 11-15 January 2007 caused by strong easterly winds (Fig. 4).

Previous observational studies suggested the reversals of the slope current are often associated with upwelling events (Nikolopoulos et al. 2009; Pickart et al. 2009b; Schulze and Pickart 2012). The several salty bottom-intensified spikes in the time series of salinity (Fig. 13b) are due to upwelling events. Strong easterly winds along the northern Alaska coast associated with north-moving Aleutian lows are favorable for upwelling. During autumn without ice cover, upwelling can be very strong (Pickart et al. 2009b). Pickart et al. (2009b) observed a strong upwelling event with a reversal of the slope current on 6 November 2002 (Fig. 14a). In the CIOM simulation, a strong upwelling occurred on 10 October 2006, with similar current and

temperature-salinity structures (Fig. 14b). In this case, the boundary current was flowing > 1 m/s to the west as a surface-intensified jet. Coincident with the reversal jet, the isohalines and isothermals were sloped strongly upward toward the boundary, and the warm and salty Atlantic water was upwelled to the shelf.

During winter, ice cover significantly reduces the stress on the water, and thus also offshore water transport. Although upwelling events still occur, the strength is greatly reduced as shown in Fig. 14c, which is in good agreement with Pickart et al.'s (2009b) argument that the presence of pack ice might weaken the strength of the upwelling events, which is based on moored data collected during the autumn and early winter of 2002. For example, Fig. 14c shows an upwelling event that occurred around 13 January 2007. In this case, the slope current reversed to the west with the zonal velocity of about $-30 \sim -40$ cm/s (Fig. 14c). The warm and salty Atlantic water was not upwelled to the shelf.

Compared to the upwelling, downwelling events draw much less attention though they are important in water and material exchanges near the shelf break. During winter 2006/2007, there were two cyclones that caused obvious downwelling along the northern Alaska coast: one was during 25-28 February (Fig. 14d), the other during 26-28 March 2007 (not shown). The downwelling events are often associated with an enhanced slope current. Especially during 26-28 March, a very strong Arctic born cyclone covered almost the entire Arctic resulting in strong westerly winds along the northern Alaska coast which accelerated the slope current (Fig. 13a) and led to a strong downwelling event along the shelf break.

5.4 Surface Heat flux

It is well known that ice cover usually cuts off most of air-sea heat exchanges, especially the sensible and latent heat fluxes and solar radiation, while leads (open water areas in sea ice)

are remained as the major area of surface heat loss. As shown in Fig. 15, the calculated net surface heat flux (NSHF) over water in February 2007 ranges from 200 to 500 Wm^{-2} , increasing from south to north mainly due to the colder air temperatures in north. The NSHF over ice is about 1 tenth of what is simulated over water, ranging from 5 to 50 Wm^{-2} . If there is 10% open water in a grid cell, heat loss over the open water will be 20-50 Wm^{-2} , which is more than twice the heat loss over thick ice in the northern portion of the domain, and at least comparable to the heat loss over thin ice in the southern portion of the domain.

Heat loss over first-year ice is larger than over multi-year ice (Fig. 15b, Fig. 16a). The green lines represent the 15% ice edge of the multi-year ice cover as inferred from the ice cover minima during the previous summer, following the method used by Comiso (2012). Multiyear ice cover dominates the northern Beaufort Sea, while first-year ice cover dominates the Chukchi Sea and the southern Beaufort Sea. The winter mean heat loss is about 15 Wm^{-2} over multi-year ice in the Beaufort Sea basin, while it was about 33 Wm^{-2} over first-year ice in the Chukchi Sea (Table 2 and Fig.16a) which compares reasonably well to previous studies. For example, Maykut (1978) gave a climatological estimation of 13.5 Wm^{-2} heat loss over multi-year ice (winter mean based on his Table 3). Persson et al. (2002) provided an estimation of 17 Wm^{-2} over pack ice based on direct observations during winter 1997/1998 (winter mean based on his Figure 21). Among the three components of the surface heat flux, the net long-wave radiation dominates heat exchange over both thin and thick ice, while sensible and latent fluxes are of secondary importance (Table 2, Fig. 16b). Further, over the thick ice, the latent heat flux is often smaller than the sensible heat flux (Table 2).

Storms have a large influence on heat loss from ice and water surfaces due to strong winds and large temperature and moisture differences. Figures 17 shows NSHF in three stages of a

storm which occurred on 25-27 February 2007. In the developing stage, at 18 GMT February 25 (Fig. 17a), just before merging, the cyclone was still weak with two centers. The heat loss is smaller than 60 Wm^{-2} in most of the domain, and it is larger in the outer area than the inner area of the cyclone, which is consistent with the distribution of wind speed associated with cyclones. At the peak of the storm at 12 GMT 27 February (Fig. 17b), the low pressure center was located in the northern portion of the domain with strong westerly winds in the southern portion of the cyclone in the Chukchi-Beaufort Sea, resulting in the largest heat loss ($80\text{-}140 \text{ Wm}^{-2}$). In the decay stage at 18 GMT 28 February (Fig. 17c), the cyclone weakened and left the domain, leading to weak winds and smaller heat loss.

The time series of domain-averaged total NSHF clearly show storm impacts with peaks occurring in peaks of the mentioned storms. During the peak of several storms, heat loss is 1.5-2 times that during normal days (Fig. 18a). This implies that recent increased storm activities in the Arctic will lead to more heat loss to the atmosphere. The increase of heat loss is mainly from leads via sensible heat flux (Fig. 18b), and heat loss over ice surface is quite stable ($\sim 20 \text{ Wm}^{-2}$) with some small fluctuations in response to the storms (Fig. 18a). During several storms, heat loss over ice decreased (blue line in Fig. 18a), because the storms brought warmer air from the south (Figs. 2 and 4). The most obvious event occurred on 2-3 February 2007, when a very strong northward-moving Aleutian low brought Pacific warm air, and the area-averaged surface air temperature was about -10°C . During 2-3 February, heat loss over open water was also small (Fig. 18a).

Sensible heat flux over open water contributes the most to the increasing heat loss during storms, and net long wave radiation and latent heat flux have smaller fluctuations (Fig. 18c). The open water temperature is close to freezing point (about -2°C for sea water), while the surface

air temperature ranges from -10° to -35°C (Fig. 2). The large air-sea temperature difference results in a very large sensible heat flux, which increases dramatically during storms due to strong winds. Net heat loss over open water increased from around 300 Wm^{-2} in normal days to about $700\text{--}800\text{ Wm}^{-2}$ in peaks of several storms (Fig. 18c). The increase in sensible heat flux is the largest among the three components, which increases from about 100 to 400 Wm^{-2} . The latent heat flux increases from about 50 Wm^{-2} to about 150 Wm^{-2} , which is also significant relative to its value. On the contrary, the net long wave radiation shows a slight decrease during the storms, as the warm air emitted more downward long-wave radiation.

6. Conclusions

A Coupled Ice-Ocean model (CIOM) was used to explore storm impacts on the surface heat flux, sea ice, and oceanic dynamics in the Chukchi-Beaufort Sea. During winter 2006/2007, there were six main storms affecting the study domain; four were north-moving Aleutian lows, and two were born or intensified in the Arctic. Six-hourly NCEP2 reanalysis forcing was used to force the CIOM. The main conclusions from the model analysis are as follows:

(1) Driven by the 6-hourly forcing, the CIOM can reproduce the storm impacts successfully. Simulated sea ice movements are comparable to the satellite observations. Strong winds associated with storms result in fast moving ice, which is faster than water because the ice cover significantly reduced the momentum exchange between the atmosphere and water. The gap between sea ice and surface water speed increases with the wind speed. The angle between the wind and sea ice velocity varies with wind speed. When the wind is weak ($< 5\text{ m/s}$), the angle is scattered due to the weak wind effects. When the wind becomes stronger, the angle tends to around $20\text{--}40^{\circ}$ to the right of the wind.

(2) Storms can alter the shelf circulation in days. The most obvious response is the pathways of the Pacific inflow. Under the forcing of easterly winds, the Pacific water flows northward directly into the interior basin. Under westerly winds, most of the Pacific inflow goes no further than the latitude of Wrangel Island. During a winter with heavy ice cover, both upwelling and the associated reversal of the slope current along the shelf break induced by the storms are reasonably reproduced. Compared to autumn, the strength of upwelling events is greatly reduced.

(3) Storms have significant influence on surface heat flux due to strong winds and large temperature differences. During the peak of several storms, heat loss to the atmosphere was about twice that during non-storm days. The increase in heat loss is mainly attributed to sensible and latent heat fluxes over the open water. Heat loss over ice was quite stable with some small fluctuations responding to the storms.

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Appendix A. Sea ice model description

The coupled ice-ocean model was described in great detail in Wang et al. (2002) which described the model development and results from a pan-Arctic simulation. In the following, we only describe the parts necessary for the current paper.

The ice velocity $\bar{V}_i$ is determined from the momentum equation:

$$m \frac{d\bar{V}_i}{dt} + m f \bar{k} \times \bar{V}_i = -mg \nabla H + \bar{\tau}_{ai} - \bar{\tau}_{wi} + \bar{F} \quad (1)$$

where f is the Coriolis parameter, m is the ice mass in a grid cell, ∇H is the gradient of sea surface elevation, $\bar{F}$ is the internal stresses (Hibler 1979), and $\bar{\tau}_a$ and $\bar{\tau}_w$ are the air and water stresses, respectively. They are determined by the bulk formulas:

$$\bar{\tau}_{ai} = \rho_a C_a \left| \bar{V}_a \right| \bar{V}_a \quad (2)$$

$$\bar{\tau}_{wi} = \rho_w C_w \left| \bar{V}_w - \bar{V}_i \right| (\bar{V}_w - \bar{V}_i) \quad (3)$$

where $\bar{V}_a$ is the wind velocity vector, $\bar{V}_w$ is the current velocity vector of the upmost ocean layer, C_a ($=1.2 \times 10^{-3}$) and C_w ($=5.5 \times 10^{-3}$) are the bulk coefficients of wind and water stress respectively and ρ_a ($=1.3 \text{ kg m}^{-3}$) and ρ_w ($=1.025 \times 10^3 \text{ kg m}^{-3}$) are the air and the seawater density respectively.

When a cell is covered with ice at concentration A , the net stress at the water surface is

$$\bar{\tau}_w = -\bar{\tau}_{wi} A + \bar{\tau}_{aw} (1 - A) \quad (4)$$

where the air-water stress is given in a similar way to $\bar{\tau}_{ai}$:

$$\bar{\tau}_{aw} = \rho_a C_a \left| \bar{V}_a \right| \bar{V}_a \quad (5)$$

The heat budget is calculated at the ice or open water surface and is a balance of the shortwave (I_0) and longwave radiation, the sensible and latent heat flux, and the conductive heat

through the ice column. The surface ice temperature is determined from the surface heat balance equation:

$$I_0(1-\alpha) + H_{lw} + H_s + H_l - F_c = 0 \quad (6)$$

where the sensible heat flux H_s , latent heat flux H_l , and net long wave radiation are calculated using:

$$H_s = \rho_a C_s C_p |V_a| (T_a - T_{sfc}) \quad (7)$$

$$H_l = \rho_a C_l l_e |V_a| (q_a - q_{sfc}) \quad (8)$$

$$H_{lw} = -\epsilon \delta \left[(1 - k_c C) (0.254 - 4.95 \times 10^{-5} e_a) T_a + 4(T_{sfc} - T_a) \right] T_a^3 \quad (9)$$

$$F_c = -k_i (T_{sfc} - T_f) / h_i \quad (10)$$

where C_p ($=1005 \text{ J kg}^{-1} \text{ K}^{-1}$) is the specific heat of air, l_e ($=2.5 \times 10^6 \text{ J/kg}$) is the latent heat of evaporation, T_a and T_{sfc} are the surface air and water or ice temperature respectively, and q_a and q_{sfc} are the specific humidity of the surface air and water (or ice) surface, respectively. The heat transfer coefficients, C_s and C_l , depend on wind speed and are set based on Large and Pond (1982). For example, at a given wind of 5 m/s, the C_s is 1.06×10^{-3} (0.6×10^{-3}) for an unstable (stable) condition, and C_l is 1.1×10^{-3} . ϵ ($=0.95$) is the emissivity of sea surface relative to a black body. δ ($=5.67 \times 10^{-8}$) is Stefan-Boltzmann constant, k_c ($=0.7$) is the cloud factor, C is the cloud fraction, T_a in eq. (9) is the air temperature in Kelvin, K_i is the thermal conductive coefficient, and T_f is the freezing temperature of seawater on the bottom ice surface.

q_a , and q_{sfc} are calculated by

$$q_a = \frac{\epsilon e}{p - (1 - \epsilon) e} \quad (11)$$

$$q_{sfc} = \frac{\epsilon e_s}{p - (1 - \epsilon) e_s} \quad (12)$$

where $\epsilon=0.622$, p is air pressure (~ 1013 hPa), water vapor pressure, e , and saturated water vapor pressure, e_s , are determined by the following empirical formula

$$e_s = 6.11 \cdot 10^{a(T_{sfc}-273.16)/(T_{sfc}-b)} \quad (13)$$

where $a=17.26$, $b=35.86$ at water surface, while $a=21.87$, $b=7.66$ at ice surface. Using surface dew point temperature, T_d , to replace T_{sfc} in (13), water vapor pressure, e , can be obtained.

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Table 1 List of the six cyclones that occurred during winter 2006/2007

No .	Date	Time of Peak Wind Speed	Central Pressure (hPa)	Origin Location
1	17-21 Dec. 2006	18Z, 20 Dec. 2006	956	Bering Sea/Gulf of Alaska
2	10-15 Jan. 2007	00Z, 13 Jan. 2007	968	Western Bering Sea
3	31 Jan.-1 Feb. 2007	12Z, 31 Jan. 2007	964	Aleutian Island
4	2-3 Feb. 2007	12Z, 2 Feb. 2007	972	Aleutian Island
5	25-28 Feb. 2007	18Z, 26 Feb. 2007	1000	Great Slave Lake and Sea of Okhotsk
6	26-28 Mar. 2007	00Z, 27 Mar. 2007	964	NE of the Greenland Sea

Table 2 CIOM simulated DJFM mean surface heat fluxes (Wm^{-2}) during winter 2006/07

Type	NSHF ^a	Hs ^a	HI ^a	HIw ^a
Open water	361.2	205.5	68.2	108.6
First-year ice	33.0	5.8	-1.3	30.3
Multi-year ice	15.1	-5.2	-2.1	24.0

^a NSHF-Net surface heat Flux; Hs-Sensible heat flux;
HI-Latent heat flux; HIw-net long wave Radiation

Figure Captions

Figure 1 A schematic diagram for coastal circulation in the Chukchi-Beaufort Seas: Alaskan

Coast Current with the origin of freshwater (light blue); Central Branch Current (medium blue); Herald Canyon Branch (dark blue); Slope Current (red); Beaufort Gyre (purple); East Siberian Current (green). Depths are in meters (Courtesy of Tom Weingartner). The mooring locations of the SBI array are indicated (yellow).

Figure 2 Domain-averaged (66.5°N to 75°N and 176°E - 125°W) 6 hourly 2 m air temperature ($^{\circ}\text{C}$; dashed line) and surface wind speed (m/s; solid line) during winter 2006/2007 (December-March) (From NCEP reanalysis 2). Letters “C” indicates the six storm events listed in Table 1.

Figure 3 Daily trajectories of the selected cyclones. The trajectories are based on the 6 hourly NCEP2 SLP data. Unless noted, the solid dots in the trajectories represent the position of cyclone center at 18 GMT each day.

Figure 4 Snapshots of sea level pressure (black contours, interval: 4hPa), 2m air temperature (shaded, $^{\circ}\text{C}$), and surface winds (vectors, m/s) for the six selected storms.

Figure 5 (a) NSIDC observed (black line) and CIOM simulated (green line) daily sea ice area (ice concentration $> 15\%$) in the Chukchi and Beaufort Sea from 1 October 2006 to 1 October 2007; (b) CIOM simulated sea ice concentration in February 2007.

Figure 6 NSIDC observed (black line) and CIOM simulated (red line) daily domain-averaged sea ice speed.

Figure 7 (a) NSIDC observed and (b) CIOM modeled sea ice movements (cm/s) on 19 December 2006. (c) and (d), same as (a) and (b), but for 27 March 2007.

Figure 8 CIOM simulated DJFM mean upmost 50 m ocean currents (black vectors, m/s). Surface winds (blue vectors, m/s).

Figure 9 (a) Left: Simulated vertical cross-sections of DJFM mean zonal velocity (cm/s) along 156.5 °W. Contours show easterly currents (red) and westerly currents (blue) in cm/s. Right: temperature (shaded, °C) and salinity (contours) along 156.5 °W. (b) Same as (a), but for the section along the direction of the mooring array. The x-axis is cross-slope distance in reference to BS1 station, consistent with Nikolopoulos et al (2009). (c) Left: Observed winter mean vertical section of along stream velocity (cm/s). Right: Potential temperature (color shaded, °C) and salinity (contours, right panel) from the mooring array near 152°W averaged over Oct. 2002-Feb. 2003 (from Nikolopoulos et al. 2009).

Figure 10. (a) Averaged ice speed (red line, cm/s), surface water speed (blue line, cm/s), and 10-m wind speed (black line, m/s) for central Chukchi Sea; (b) Same as (a), but for central Beaufort Sea; (c) The plane scatterplot between sea ice and water speed difference ($U_i - U_w$) and wind speed in the central Chukchi Sea for the period from 1 January to 31 March 2007; (d) Same as (c), but for angle between ice and wind.

Figure 11. Top 50m water velocity (black vectors, m/s) and 10-m winds (green vectors, m/s) averaged over the storms that occurred on (a) December 17-21, 2006, (b) January 10-15, 2007, and (c) March 26-28, 2007.

Figure 12 Time series of (a) 10m winds (m/s), (b) water zonal velocity (cm/s) at 156.5°W, 71.35°N from December 1, 2006 to March 31, 2007

Figure 13 Time series of (a) water zonal velocity (cm/s) and (b) water salinity (PSU) at 152°W, 71.32°N in the shelf break of the Beaufort Sea from December 1, 2006 to March 31, 2007

Figure 14 (a) Observed vertical sections of velocity and potential temperature and salinity from the mooring array near 152°W, showing an upwelling event in early November 2002 (from Pickart et al. 2009); CIOM simulated vertical sections of (left) along stream velocity (cm/s) and

(right) temperature (shaded, °C) and salinity (contour) along the direction of the mooring array near 152°W for (b) 6 Oct. 2006, (c) 13 Jan. 2007, and (d) 27 Feb. 2007.

Figure 15 The CIOM simulated monthly mean net surface heat flux over (a) water and (b) ice in February 2007 (Wm^{-2}). Note the NSHF in (a) is not area-weighted. The green line indicates 15% ice edge during September 2006.

Figure 16. (a) Net surface heat flux over sea ice in the central Chukchi Sea (black line with triangle) and the central Beaufort Sea (blue line with circle); (b) Sensible (red line) and latent (green line) heat fluxes and net long wave radiation (blue line) over sea ice in the central Chukchi Sea (solid line with triangle) and central Beaufort Sea (solid line with circle). Upward is positive. Units are W/m^2 .

Figure 17. Calculated net surface heat flux (Wm^{-2} , positive upward) with wind vectors overlain during the passage of the cyclone on 25-28 February 2007.

Figure 18. (a) Domain averaged total net surface heat flux (black line with circle), net surface heat flux over water (area weighted; green line with solid circle) and sea ice (blue line with triangle); (b) Domain averaged sensible (red line with circle) and latent (green line with circle) heat fluxes and net long wave radiation (blue line with triangle) over the whole grid (both water and sea ice); (c) Domain averaged net (black), sensible (red) and latent (green) heat fluxes and net long wave radiation (blue) over open water (the fluxes are not area weighted). Upward is positive. Units are W/m^2 .

Highlights

The CIOM reasonably reproduces storm impacts on sea ice and ocean dynamics.

Storms alter pathways of the Pacific inflow water in the Chukchi Sea shelf.

Storm-induced upwelling and slope current reversal are successfully reproduced.

We examine air-ice-sea surface heat flux during the passage of storms.

We describe the evolution of storms affecting the Chukchi/Beaufort Sea.

Figure 1. A schematic diagram for coastal circulation in the Chukchi-Beaufort Seas: Alaskan Coast Current with the origin of freshwater (light blue); Central Branch Current (medium blue); Herald Canyon Branch (dark blue); Slope Current (red); Beaufort Gyre (purple); East Siberian Current (green). Depths are in meters (Courtesy of Tom Weingartner). The mooring locations of the SBI array are indicated (yellow).

Figure 2 Domain-averaged (66.5°N to 75°N and 176°E-125°W) 6 hourly 2m air temperature (°C; dashed line) and surface wind speed (m/s; solid line) during winter 2006/2007 (December-March) (From NCEP reanalysis 2). Letters “C” indicates the six storm events listed in Table 1.

Fig. 3 Daily trajectories of the selected cyclones. The trajectories are based on the 6 hourly NCEP2 SLP data. Unless noted, the solid dot in the trajectories represents the position of cyclone center at 18 GMT each day.

Figure 4. Snapshots of sea level pressure (black contours, interval: 4hPa), 2m air temperature (shaded, °C), and surface winds (vectors, m/s) for the six selected storms.

Figure 5. (a) NSIDC observed (black line) and CIOM simulated (green line) daily sea ice area (ice concentration > 15%) in the Chukchi and Beaufort Sea from 1 October 2006 to 1 October 2007; (b) CIOM simulated sea ice concentration in February 2007.

Figure 6. NSIDC observed (black line) and the CIOM simulated (red line) daily domain-averaged sea ice speed.

Figure 7. (a) NSIDC observed and (b) CIOM modeled sea ice movements (Unit: cm/s) on 19 December 2006. (c) and (d), same as (a) and (b), but for 27 March 2007.

Figure 7 (Continued)

Figure 8 CIOM simulated DJFM mean upmost 50 m ocean currents (black vectors, m/s).

Surface winds (blue vectors, m/s).

Figure 9 (a) Left: Simulated vertical cross-sections of DJFM mean zonal velocity (cm/s) along 156.5 °W. Contours show easterly currents (red) and westerly currents (blue) in cm/s. Right: temperature (shaded, °C) and salinity (contours) along 156.5 °W. (b) Same as (a), but for the section along the direction of the mooring array. The x-axis is cross-slope distance in reference to BS1 station, consistent with Nikolopoulos et al (2009). (c) Left: Observed winter mean vertical section of along stream velocity (cm/s). Right: Potential temperature (color shaded, °C) and salinity (contours, right panel) from the mooring array near 152°W averaged over Oct. 2002-Feb. 2003 (from Nikolopoulos et al. 2009).

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Fig. 12 Time series of (a) 10m winds (m/s), (b) water zonal velocity (cm/s) at the 156.5°W, 71.35°N from December 1, 2006 to March 31, 2007

Fig. 13 Time series of (a) water zonal velocity (cm/s) and (b) water salinity (PSU) at the site 152°W, 71.32°N in the shelf break of the Beaufort Sea from December 1, 2006 to March 31, 2007

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Figure 17 Calculated net surface heat flux (Wm^{-2} , positive upward) with wind vectors overlain during the passage of the cyclone on 25-28 February 2007.

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