Pacific Ocean inflow: Influence on catastrophic reduction of sea ice cover in the Arctic Ocean

Koji Shimada,[1] Takashi Kamoshida,[1] Motoyo Itoh,[1] Shigeto Nishino,[1] Eddy Carmack,[2] Fiona McLaughlin,[2] Sarah Zimmermann,[2] and Andrey Proshutinsky[3]

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[1] The spatial pattern of recent ice reduction in the Arctic Ocean is similar to the distribution of warm Pacific Summer Water (PSW) that interflows the upper portion of halocline in the southern Canada Basin. Increases in PSW temperature in the basin are also well-correlated with the onset of sea-ice reduction that began in the late 1990s. However, increases in PSW temperature in the basin do not correlate with the temperature of upstream source water in the northeastern Bering Sea, suggesting that there is another mechanism which controls these concurrent changes in ice cover and upper ocean temperature. We propose a feedback mechanism whereby the delayed sea-ice formation in early winter, which began in 1997/1998, reduced internal ice stresses and thus allowed a more efficient coupling of anticyclonic wind forcing to the upper ocean. This, in turn, increased the flux of warm PSW into the basin and caused the catastrophic changes. **Citation:** Shimada, K., T. Kamoshida, M. Itoh, S. Nishino, E. Carmack, F. A. McLaughlin, S. Zimmermann, and A. Proshutinsky (2006), Pacific Ocean inflow: Influence on catastrophic reduction of sea ice cover in the Arctic Ocean, *Geophys. Res. Lett.*, *33*, L08605, doi:10.1029/2005GL025624.

1. Introduction

[2] Sea-ice reduction have been recognized as an indicator of Arctic climate change [*Arctic Climate Impact Assessment*, 2004]. Variability in sea-ice extent, thickness and motion has often been correlated with atmospheric indices such as the Arctic oscillation (AO) and North Atlantic oscillation (NAO) [*Kwok*, 2000; *Polyakov and Johnson*, 2000; *Rigor et al.*, 2002; *Serreze et al.*, 2003]. Such atmospheric variability affects the frequency of cyclones entering the Arctic Ocean, and the pathways and volume flux of sea-ice outflow through Fram Strait [*Proshutinsky and Johnson*, 1997; *Kwok and Rothrock*, 1999; *Pfirman et al.*, 2004]. Prior to the mid-1990s, sea-ice reduction correlated with an increasing (positive) AO index. Since the mid-1990s, however, the AO index has decreased while the reduction of sea ice, especially in the Pacific sector of the Arctic Ocean, has not only continued but accelerated. This suggests that variability of large-scale atmospheric forcing only partially explains recent changes in the ice cover. Figure 1 shows the sea-ice concentration anomaly: September 1998–2003 mean minus September 1979–1997 mean. The area of anomalous sea-ice reduction corresponds to the area where warm Pacific Summer Water (PSW: salinity range $31.0 < S < 32.0$) is observed just beneath the surface mixed layer (20–60m) [*Coachman et al.*, 1975, *Shimada et al.*, 2001]. Overlapping sea-ice reduction and PSW distribution domains suggest a causal link to Pacific inflow through Bering Strait. Here we analyze temperature and salinity data from: SCICEX 1993–2000; Fisheries and Oceans Canada's Institute of Ocean Sciences (IOS) 1992–2002; Japan Agency for Marine-Earth Science and Technology (JAMSTEC) 1992–2002; and a joint JAMSTEC, IOS and Woods Hole Oceanographic Institution project 2003–2004. We analyze sea-ice concentration data generated using NASA team algorithm from NSIDC [*Cavalieri et al.*, 1990] and sea-ice motion data calculated using maximum cross correlation techniques from NSIDC [*Fowler*, 2003].

2. Changes in the Sea Ice Cover

[3] Figure 2a shows the time series of sea-ice concentration in September adjacent to the Northwind Ridge in the Canada Basin (73°–77°N, 150°–165°W) where maximum sea-ice reduction is observed, and illustrates that sea-ice cover reduced rapidly after 1997/1998. Before 1997, sea-ice concentration here gradually decreased at approximately −0.6% per year, similar to the rate of sea-ice decrease observed across the entire Arctic Ocean [*Maslanik et al.*, 1996]. From 1996 to 1998, however, there was a substantial decrease in late summer sea-ice concentration in the Canada Basin [cf. *Maslanik et al.*, 1999]. Figure 2a shows that sea-ice concentration decreased from 60–80% to 15–30% in the three-year running mean and that it has not subsequently recovered to pre-1997 levels. This abrupt decrease in late 1990s and absence of rebound suggests there has been a catastrophic change, a change that is not reflected in either the wintertime AO index or the local wind field in the southern Canada Basin (Figure 2b). Such divergence suggests that the reduction is not fully explained by variability in large-scale atmospheric forcing. *Overland and Wang* [2005] suggested that recent warming of surface air temperature in spring, related to changes in sea level pressure, contributed to recent reductions in sea-ice cover in the western Arctic Ocean after 2000. Here we present an additional mechanism.

3. Changes in the Upper Ocean Temperature

[4] In the Canada Basin the primary oceanic heat source to the sea-ice cover and atmosphere is not the Atlantic Water

[1]Institute of Observational Research for Global Change, Japan Agency for Marine-Earth Science and Technology, Yokosuka, Japan.
[2]Fisheries and Oceans Canada, Institute of Ocean Sciences, Sidney, British Columbia, Canada.
[3]Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, USA.

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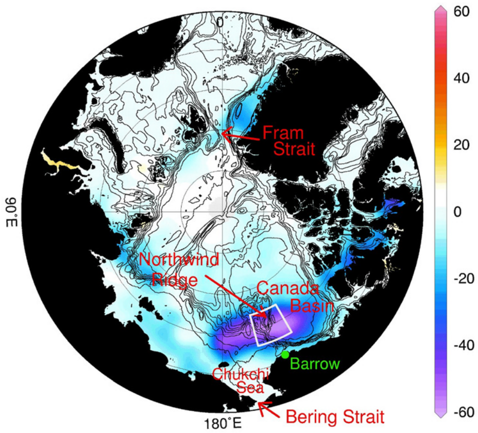

Figure 1. Sea ice concentration anomaly (%): September 1998–2003 mean minus 1979–1997 mean. Contours show bathymetry. The area outlined by the white box shows the area for which the time series of sea ice concentration and PSW temperature, shown in Figure 2 and Figure 3, were calculated.

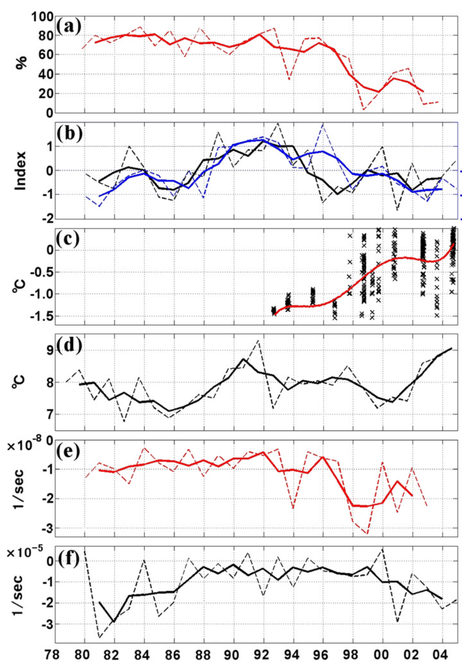

Figure 2. Time series of (a) sea ice concentration in September, 73°–77°N, 150°–165°W; (b) Arctic Oscillation index, November to January shown in black (index is provided by the Climate Prediction Center), the blue line shows eastward wind in the Canada Basin east of the Northwind Ridge: 74°–75°N, 130°–160°W; (c) potential temperature on S = 31.3, 73°–77°N, 150°–165°W (white box in Figure 1); (d) sea surface temperature in the northeastern Bering shelf, 62°–66°N, 164–170°W; (e) curl of sea ice motion ($\nabla \times \vec{u}_{ice}$) in the Canada Basin east of the Northwind Ridge, 74°–75°N, 130°–160°W; (f) curl of wind in the same area as for (e); solid curves denote three-year running mean values and dashed curves are annual values for Figures 2a, 2b, 2d, 2e, and 2f; the solid red curve in Figure 2c is a fifth degree polynomial fit.

which lies beneath the halocline but, instead, the PSW that occupies the upper portion of the halocline in the salinity range 31.0 < S < 32.0. *Shimada et al.* [2001] showed that the main pathway of PSW from the shelf region into the Canada Basin lies along the Northwind Ridge, and coincides with the location of observed maximum sea-ice reduction. They also calculated that the stored heat within the PSW layer, determined by comparing the winter and late summer temperature profiles in 1998 (just before the catastrophic change), was about 140 MJm^2. This heat has the potential to melt about 50 cm of sea ice. Figure 3 shows the time series of PSW temperature and salinity, averaged in the same area as defined in Figure 1, and including all available data collected after 1992. In the early 1990s, the maximum temperature of PSW was about −1°C, close to that of the climatological mean from the *Environmental Working Group (EWG)* [1997]. The decadal mean values from the 1950s to the 1980s were also near −1 °C. This indicates that the temperature of PSW in this region was in near steady-state until the late 1990s. By 1998, however, warming was evident as shown in the temperature time series near S∼31, in the core of PSW (Figure 2c), and the timing of this warming was well correlated with the catastrophic change in sea-ice concentration (Figure 2a). After 1998, temperature of Pacific Winter Water (PWW, S∼33) decreased concurrently with the abrupt warming of PSW. This evidence shows that the recent warming of Atlantic Water did not contribute to the recent sea-ice reduction in the Pacific sector of the Arctic Ocean. Instead, the upward heat flux from the PSW likely retarded sea-ice formation in the Canada Basin. This imbalance between sea-ice growth in winter and melt in summer may have also contributed to the reduction in multi-year ice entering and leaving this region after the late 1990s [*Rigor and Wallace*, 2004]. Is this warming trend caused by upstream changes? The time series of sea surface temperature (Figure 2d) in the Eastern Bering Sea (62°–66°N and 164°–170°W) shows that upstream temperatures, in fact, decreased in the late 1990s. Cooling in upstream source waters is corroborated by mooring data collected in Bering Strait [*Woodgate et al.*, 2005]. This discrepancy raises the additional question as to why the temperature of PSW in the western Canada Basin increased in late 1990s whereas the temperature of upstream water decreased.

4. Changes in the Sea Ice Motion

[5] If there is no external forcing except buoyancy, then warm PSW flows eastward along the shelf slope and the

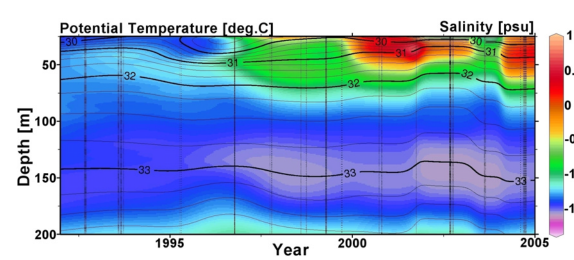

Figure 3. Time series of temperature (color) and salinity (contours) in the white box of 73°–77°N, 150°–165°W (see Figure 1).

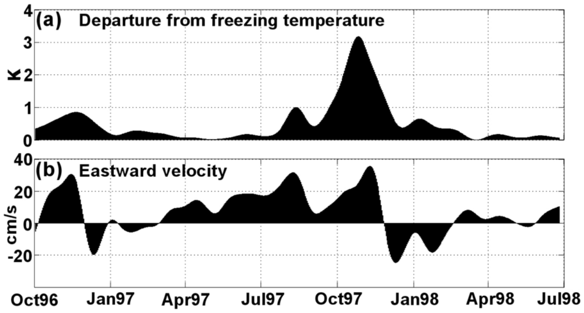

Figure 4. Time series of low-passed (>50 days) (a) departure from freezing temperature, (b) eastward velocity at 78m deep within PSW layer at the mouth of the Barrow Canyon during 1996–1998 (71°45.74′N, 155°13.57′W, bottom depth: 239m).

heat content of PSW is not effectively delivered to the central Arctic Ocean. However, the spatial distribution of the warm PSW is centered on the Northwind Ridge [*Shimada et al.*, 2001], suggesting that wind-driven, sea-ice motion (anticyclonic) overcomes the buoyancy-driven coastal current (cyclonic) and thus transports PSW heat into the central basin. Hence we examine changes in the strength of the sea-ice motion, focusing on the 1997/1998 change. Temperature time-series data from moorings at the mouth of the Barrow Canyon (71°45.74′N, 155°13.57′W) show that maximum annual PSW temperatures occur between October and November (Figure 4a). Just after the arrival of the annual maximum in PSW temperature, the direction of flow in the depth range of PSW changed from eastward to westward (Figure 4b), thus following the direction of the wind-driven Beaufort Gyre [*Proshutinsky and Johnson*, 1997]. In fact, on the Northwind Ridge the warmest temperature of PSW in the year-long record was observed during winter [cf. *Shimada et al.*, 2001]. Therefore the strength of the wind/ice-driven circulation in winter is critical to changes in PSW transport onto the Northwind Ridge. The time series of the curl of sea-ice motion ($\nabla \times \vec{u}_{ice}$) in the zonal band 74–75°N, east of the Northwind Ridge, indicates the strength of upper ocean circulation during winter (Figure 2e). In the late 1990s $\nabla\times \vec{u}_{ice}$ became more anticyclonic, by a factor of two; in contrast the wind stress curl did not increase (Figure 2f). The timing of this increased anticyclonic sea-ice motion during winter of 1997/1998 coincided with the decrease in sea-ice concentration (Figure 2a). However, this shift in sea-ice motion does not correlate to the AO index (Figure 2b). In the 1980s $\nabla \times \vec{u}_{ice}$ remained relatively constant despite changes in wind forcing, whereas after the mid-1990s sea-ice motion changed coherently with wind forcing. This non-linear response of $\nabla \times \vec{u}_{ice}$ to the wind shows that upper ocean circulation is not governed solely by wind forcing, but is also modulated by another mechanism.

5. Discussion and Perspective

[6] The coherence among sea-ice concentration, PSW temperature, and $\nabla \times \vec{u}_{ice}$ changes, both before and after 1998, imply modulation by a positive feedback mechanism (Figure 5) that involves the efficiency of penetration by

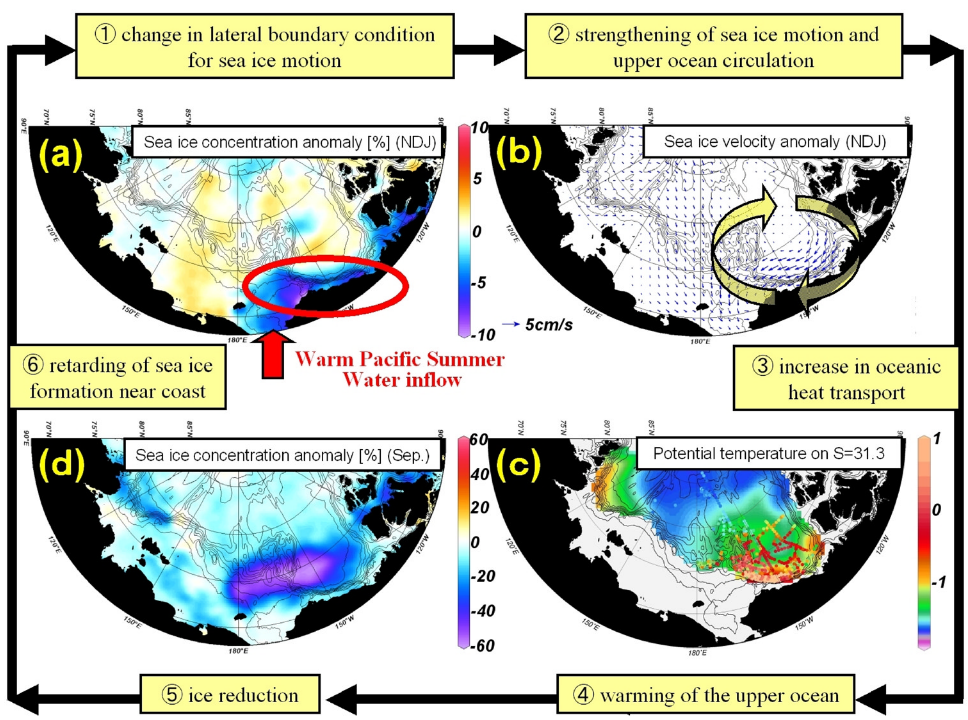

Figure 5. Feedback system. (a) Sea ice concentration anomaly for November through January [(1997 Nov.∼2003 Jan.) - (1979 Nov.∼1997 Jan.)] (b) Sea ice velocity anomaly for November through January [(1997 Nov.∼2003 Jan.) - (1979 Nov.∼1997 Jan.)] (c) Potential temperature on S = 31.3. Background color is climatology from EWG Arctic Ocean Atlas and dotted circles are from 1998–2004. (d) Sea ice concentration anomaly for September [(1998∼2003)-(1979∼1997)].

wind forcing into the ocean through the sea-ice cover which, in turn, depends on the sea-ice cover itself [cf. *Kimura and Wakatsuchi*, 2000]. If the momentum flux from the atmosphere is shielded by large internal ice stresses arising from the near non-slip boundary condition along the Alaskan coast, then the ice-driven upper ocean circulation - which acts to retroflex PSW toward the west - is insufficient to overcome the eastward, buoyancy-driven boundary current [cf. *Pickart*, 2004]. However, if sea-ice cover is reduced along the coast, then the momentum of wind is more efficiently transferred to the ice and underlying waters. Therefore the delayed development of sea-ice cover in winter enhances the retroflexion (westward turning) of PSW, just as the warmest pulse of PSW arrives on the Beaufort Slope. This anomalous heat flux into the western Canada Basin retards sea-ice formation during winter which, in turn, causes an imbalance between ice growth in winter and ice melt in summer, further accelerating sea-ice reduction. This demonstrates a unique aspect of seasonally ice-covered seas in that wind forcing is not directly delivered to the ocean but is modulated by ice cover. The potential of shifting from no-slip to free-slip boundary conditions constraining sea-ice motion implies that thresholds are established that, once crossed by an initial reduction in sea-ice, result in catastrophic change. This feedback condition is unique to ice-covered seas; in mid-latitude oceans the transport of heat in the upper ocean is mainly due to western boundary currents, such as the Kuroshio and Gulf Stream, and these will not show discontinuous or abrupt change in coupling efficiency.

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E. Carmack, F. McLaughlin, and S. Zimmermann, Fisheries and Oceans Canada, Institute of Ocean Sciences, 9860 W. Saanich Rd., Sidney, BC, Canada V8L 4B2.

T. Kamoshida, M. Itoh, S. Nishino, and K. Shimada, Institute of Observational Research for Global Change, Japan Agency for Marine-Earth Science and Technology, Yokosuka 237-0061, Japan. (shimadak@jamstec.go.jp)

A. Proshutinsky, Woods Hole Oceanographic Institution, MS 29, 360 Woods Hole Road, Woods Hole, MA 02543, USA.