Connections among ice, runoff and atmospheric forcing in the Beaufort Gyre

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Abstract. During SHEBA, thin ice and freshening of the Arctic Ocean surface in the Beaufort Sea led to speculation that perennial sea ice was disappearing [McPhee et al., 1998]. Since 1987, we have collected salinity, δ18O and Ba profiles near the initial SHEBA site and, in 1997, we ran a section out to SHEBA. Resolving fresh water into runoff and ice melt, we found a large background of Mackenzie River water with exceptional amounts in 1997 explaining much of the freshening at SHEBA. Ice melt went through a dramatic 4–6 m jump in the early 1990s coinciding with the atmospheric pressure field and sea-ice circulation becoming more cyclonic. The increase in sea-ice melt appears to be a thermal and mechanical response to a circulation regime shift. Should atmospheric circulation revert to the more anticyclonic mode, ice conditions can also be expected to revert although not necessarily to previous conditions.

1. Introduction

The loss of Arctic Ocean ice stands out as a pivotal change that would affect the global heat balance [Curry et al., 1995] and have unthinkable consequences for arctic inhabitants. Therefore, considerable attention has focused on recent changes in the ice including thinning, melting, reduced area, modified circulation and shifts in the onset of seasons [McLaren et al., 1990; Chapman and Walsh, 1993; Johannessen et al., 1995; Serreze et al., 1995; Maslanik et al., 1996; Cavalieri et al., 1997; Smith, 1998; Yueh and Kwok, 1998]. Arctic sea ice undergoes change which may be exported to the Atlantic where it could stall global thermohaline circulation [Aagaard and Carmack, 1989]. In 1989, a shift in the atmospheric pressure field was manifested in the Beaufort Sea by increased cyclonicity, enhanced ice divergence and a decrease in the size of the Beaufort gyre [Serreze et al., 1989; Walsh et al., 1996; Maslanik et al., 1996; Proshutinsky and Johnson, 1997; Thompson and Wallace, 1998]. This change in ice regime was all too apparent in October 1997, when ice of adequate thickness could not be found and SHEBA was installed on uncomfortably thin ice [McPhee et al., 1998; Welch, 1998].

Ice climate change is difficult to detect despite formidable observing instruments (cf. Parkinson [1991]; Wadhams [1997]). The change can appear in several forms (e.g., ice area, ice thickness, season length); careful time series are difficult to collect; ice extent has an enormous annual cycle; and ice varies from seasonal near the margins to multiyear in the interior.

Ice occupies three dimensions whereas observations are often limited to two dimensions so that the cause of an observed change may remain equivocal. An appealing way to measure change in the ice is to seek its imprint in the upper ocean [McPhee et al., 1998]. More ice production means more salt and, conversely, more melting means more fresh water. The ocean integrates the net volume of fresh water but caution is required particularly on two points. First, the ice goes through a dramatic seasonal cycle (~2 m in first-year ice) and one must recognize this source of variance in comparing years. Second, resolution into runoff and ice melt components is a prerequisite to understanding freshwater balance.

Between 1987 and 1997 we collected bottle data at six intervals in the Canada Basin (Figure 1). Here, we will use sectional data (1997) to describe spatial variation in runoff and ice melt, present freshwater budgets, and relate these to the circulation change that occurred in 1989.

2. Data and Methods

Samples for salinity and δ18O were collected at Station A (Figure 1) in 1987, 1989, 1990, 1991 (Stn L144), 1995 (section) and 1997 (section). Sampling took place after the annual melt reached its maximum and during a period when ice melt changes little with time (Figure 2). Because late September follows a period dominated by cyclones, the pack is at its most open and clockwise circulation may be stalled or even reversed [Serreze et al., 1989].

Analytical methods have been fully described elsewhere [Guay and Falkner, 1997; Macdonald et al., 1995]. Salinity was determined with a precision of better than ±0.02; oxygen isotope composition (δ18O), referenced to V-SMOW, was measured to a precision of < 0.07‰; barium as determined on unacidified samples, to a precision of ±2%.
Resolving fresh water into runoff and sea-ice melt (SIM) components using δ18O and salinity measurements is well described for the Beaufort Sea [Macdonald et al., 1995]. Assuming water to be a mixture of saline water, runoff and ice melt and choosing an appropriate composition for the three end-members, the composition of any sample can be calculated. Based on Macdonald et al. [1995] we used the following (salinity, δ18O): saline (33.1, -1.1), runoff (0.15, -20.3), ice melt (6.0, -0.43). For the saline end-member we used the Pacific halocline mode water because it provides a consistent, relatively invariant foundation for the top 200 m of the southern Beaufort Sea. The δ18O of ice was chosen by applying a fractionation of 2.6‰ from surface water at the site. The runoff component includes precipitation (≈ 20 cm yr⁻¹) and river inflow. It is crucial to note that positive and negative values are meaningful for SIM: positive values imply an excess of ice melt whereas negative values imply an excess of ice formation recorded as enhanced salt in the water.

3. Results and Discussion

In 1997, positive SIM was found only in surface water under the pack (Figure 3a, yellow contours) and there is a clear demarcation between negative and positive values at the ice edge (Stn 9). Below 20–30 m under the pack, SIM was negative implying a net production of ice. The water at 30–60 m is supplied with salt by local ice production only in winter and is, therefore, a remnant of the previous winter’s polar mixed layer (PML) [McPhee et al., 1998]. Beneath the PML where the water is not in seasonal contact with the local surface, the negative SIM component implies advection of salt-enriched water probably from shelves. In support of this, Figure 3a shows large negative SIM throughout the water near the shelf break. Over the mid-toouter

shelf ice diverges in winter and ice growth is enhanced in extensive flaw leads [Macdonald et al., 1995]. Within the pack (Stns 11–14), total SIM inventory (0–150 m) remains fairly constant at −5 to −6 m (Figure 3a, yellow bars).

Runoff exhibits maxima at Stations 3–6 and 9–12 and its inventory (0–150 m) varies widely (11–19 m) along the section (Figure 3b, blue bars). The strong coherence between runoff and Ba near the surface (Figures 3b and c) identifies the Mackenzie River as the origin [Guay and Falkner, 1997]. This conclusion is supported by satellite imagery (July 1998) showing the Mackenzie plume as a coherent filamentous structure up to 400 km off the shelf (Engle and Weingartner, priv. comm.; see cover).

Runoff provides a strong control on SIM. In Figure 3a, the large negative SIM at Stations 1–6 (0–25 m) is probably the result of ice formation over the shelf in winter followed in early spring by leakage of Mackenzie River into the flaw-lead system (Figure 3b). Fresh water from the river entrains salt as it spreads beneath sea ice and also creates stratification so that brine rejected from later sea-ice formation remains in the buoyant surface layer [Macdonald et al., 1995]. There is no evidence of ice melting at nearshore stations and we suspect that ice drifted offshore early. The outer runoff maximum (Stns 9–11) is part of the late summer Mackenzie plume which delivers heat and stability to trap solar heat near the surface. As a result, the maximum in ice melt is found at the outer edge of the plume and just within the pack where stability, heat (not shown) and ice coincide.

Assuming water from 30–60 m under the pack represents the previous winter’s PML, we can estimate change in surface freshwater content between June (i.e., when melt starts, cf. Figure 2) and October 1997 (black line in the histograms on Figures 3a and b). Stations 3–6 have increased negative SIM and runoff during this time, something that could best be explained by advection of a spreading plume enriched with brine. At the outer stations, about 1.5–4.5 m of ice have melted.  

Figure 2. Sampling times in relation to freshwater cycles. Solid line shows the change in mean ice thickness for multi-year ice at the SHEBA site Pittsburg [The SHEBA Team, 1999]; dashed line shows the average Mackenzie River flow (1972–1992: Environment Canada), and shaded bands (±sd) show the average date of spring melt and autumn freeze-up (1979–1996 [Smith, 1998]).
However, the implied melting was countered by a saltier PML and the SIM inventory (0–150 m) decreased slightly (Figure 4a, yellow bars). The big shift in SIM inventory came between 1989 and 1991 when it increased by the equivalent of 4–6 m of melt. Since then, the SIM inventory has remained fairly constant while melt has propagated into the PML and deeper, suggesting both mixing and advection. Unlike SIM, runoff shows no step but varies widely (9–18 m inventory), especially in the top 10 m, but also down to the depth of the PML (Figure 4b). Although we have no time series Ba data, we can almost certainly identify the runoff with Mackenzie River water reaching the time series station. The largest amounts of runoff on record were observed in 1997 during the SHEBA deployment and this contributes much of the observed freshening [Welch, 1998]. Assuming the water at 50 m to reflect the previous winter's PML, we have estimated the June–October inventory changes for each year (Figures 4a and b; black line in histograms). Excluding 1997, the average runoff inventory during the time series (12 m) and the average seasonal input (1.1 m) imply a freshwater residence time of about 10 years, in agreement with earlier estimates [Ostlund, 1982].

The large increase in SIM after 1989 is not related to runoff. Instead, it coincides with a change in the Arctic Oscillation Index [Overland et al., 1999], a subsequent shift toward less anticyclonic wind forcing over the Arctic Ocean [Proshutinsky and Johnson, 1997] and a record minimum in ice extent [Serreze et al., 1995].

Figure 3. The 1997 section out to the SHEBA site showing (a) sea-ice melt, (b) runoff and (c) Ba. Contours show fractional compositions for sea-ice melt (yellow positive, green negative) and runoff. The bars above panels (a) and (b) show the freshwater inventory (0–150 m) by source. Negative inventories of ice melt imply a net production of ice. The black line in the histograms shows the seasonal change (June–October) in freshwater components calculated by referencing to the remnant of last year's PML at 40–50 m.

This amount of melt can be accounted for by in situ melting plus advection of ice melt that occurred at the edge of the pack and has been transported into the pack with the Mackenzie plume. In support of this, Figure 3b (black line in the histogram) shows much seasonal runoff under the pack, particularly at Stations 9 and 11. SIM in the surface at the time series station shifted from negative in 1987 to positive in 1989 (Figure 4a).

Figure 4. Time series from 1987 to 1997 showing (a) sea-ice melt and (b) runoff. The meaning of the histograms and black lines parallel those in Figure 3.
Similar changes have occurred previously at 5–7 yr intervals and the sea ice clearly exhibits alternating wind-forced circulation regimes. The change in wind field has two important consequences for the Beaufort Gyre. In the cycloonic mode the pack ice becomes more divergent [Serreze et al., 1989] and ice circulation changes such that ice from north of the archipelago can no longer enter the gyre (Pfirman, priv. comm.). Opening the pack and shutting off the source of multi-year ice both lead to melting and thinning of ice. Furthermore, the trend toward easily melted first-year ice would lead to greater open water at the end of summer, something that has been observed since 1989 (Parkins, priv. comm.).

The 1989 regime shift is reflected by a step in SIM during the early 1990s as the ice in the gyre re-adjusted thickness and distribution. It is unreasonable that 4–6 m of ice melted throughout the region. The large change in SIM suggests that some of the melt advected to the site. The southern Beaufort Sea near the edge of the pack is probably a region of enhanced melting supported by ice advection, water stability and heat. Although divergence of pack ice would lead to thermally-forced thinning of the ice, mechanical forcing may play an even greater role. Reduced ridging during periods of divergence leads to reduced ice volume and reduced resistance to summer melting. Regardless of mechanism, thermodynamics and dynamics both operate during cycloonic periods to thin and melt the ice. The latest cycloic mode has been accompanied by exceptionally reduced ice in the Beaufort Sea; when the system reverts to the anticyclonic mode, ice conditions can be expected also to revert, but not necessarily return to previous conditions. Changes in the Arctic Oscillation forced by greenhouse gases, as proposed by Shindell et al. [1999], would be an effective way to change ice climate in Beaufort Gyre. The data presented here underscore the importance of monitoring the Canada Basin waters systematically using a suite of physical and chemical tracers.

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References


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