

Circulation of summer Pacific halocline water in the Arctic Ocean

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[1] We present an analysis of Arctic Ocean hydrographic and sea ice observations from the 1990s, with a focus on the circulation of water that originates in the North Pacific Ocean. Previous studies have shown the presence of two varieties of relatively warm “summer halocline water” in the vicinity of the Chukchi Sea, i.e., the relatively fresh Alaskan Coastal Water (ACW) and the relatively saltier summer Bering Sea Water (sBSW). Here we extend these studies by tracing the circulation of these waters downstream into the Arctic Ocean. We find that ACW is generally most evident in the southern Beaufort Gyre, while sBSW is strongest in the northern portion of the Beaufort Gyre and along the Transpolar Drift Stream. We find that this separation is most extreme during the early mid-1990s, when the Arctic Oscillation was at historically high index values. This leads us to speculate that the outflow to the North Atlantic Ocean (through the Canadian Archipelago and Fram Strait) may be similarly separated. As Arctic Oscillation index values fell during the later 1990s, ACW and sBSW began to overlap in their regions of influence. These changes are evident in the area north of Ellesmere Island, where the influence of sBSW is highly correlated, with a 3-year lag, with the Arctic Oscillation index. We also note the presence of winter Bering Sea Water (wBSW), which underlies the summer varieties. All together, this brings the number of distinct Pacific water types in our Arctic Ocean inventory to three: ACW, sBSW, and wBSW. *INDEX TERMS:* 4207

Oceanography: General: Arctic and Antarctic oceanography; 4512 Oceanography: Physical: Currents; 4536 Oceanography: Physical: Hydrography; *KEYWORDS:* Bering Sea Water, Alaskan Coastal Water, Arctic Ocean

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1. Introduction

[2] What is the circulation and interannual variability of summertime Pacific origin halocline waters in the Arctic Ocean? Here we refer to waters that enter the Arctic Ocean from the Chukchi Sea, after passing through Bering Strait from the North Pacific Ocean. A typical profile from the Canadian Basin is shown in Figure 1. A subsurface temperature maximum within a low-salinity layer (roughly, $31 < S < 33$) marks the summer Pacific halocline water, which has subducted below the fresher (i.e., lighter) surface mixed layer. Some Pacific halocline water also contains a high complement of dissolved nutrients [e.g., Jones and Anderson, 1986], which may be used to trace its circulation [Jones et al., 1998]. Coachman and Barnes [1961] were the first to describe the nature of the shallow temperature maximum within the Arctic Ocean, following earlier studies by several Russian researchers. Areas that lack this layer (i.e., the Eurasian Basin of the Arctic Ocean) generally retain near-freezing surface temperatures down to 100-m depth or more, even as the salinity increases rapidly. This is

known as a cold halocline layer. In contrast, Figure 1 shows that the Canadian Basin contains what Steele and Boyd [1998] referred to as the “cool halocline.”

[3] Why do we care about the fate of Pacific water downstream of the Chukchi Sea? The relatively fresh Pacific water ($S < 33.5$) comprises about two thirds of the Canadian Basin halocline by thickness (Figure 1) and about half by freshwater content [e.g., Aagaard and Carmack, 1989]. Some component of this water carries high nutrient content throughout the Arctic Ocean and even downstream into, for example, Baffin Bay [Tremblay et al., 2002]. Finally, the freshwater flux from the North Pacific Ocean into the North Atlantic Ocean provides a “short circuit” for the global thermohaline ocean circulation [e.g., Wijffels et al., 1992; Goosse et al., 1997].

[4] Figure 2 shows the lateral extent of summer Pacific halocline waters within the Arctic Ocean Atlas’ 40-year mean summer (July–August–September) climatology [Environmental Working Group (EWG), 1998]. The temperature maximum is generally $< -1.0^{\circ}\text{C}$, and lies at a depth of 70–100 m within a salinity range of 31–33. The downstream limit of summer Pacific influence lies roughly along the 180°E meridian (i.e., the Mendeleyev Ridge) northward until about 87°N , where it angles eastward toward Northern

Greenland. A caveat is that this climatological field is laterally smoothed and vertically sparse (see the lower depth scale in Figure 2c), which may impact the spatial characteristics of water mass positions and boundaries. Another caveat is that the temperature maximum may diffuse away while other chemical signatures of Pacific origin waters remain [Moore *et al.*, 1983; Jones *et al.*, 2003]; that is, the true extent may be larger than that shown in Figure 2.

[5] *Coachman and Barnes* [1961] referred to the shallow temperature maximum simply as Bering Sea Water, a name that is still generally in use today. However, in a later study [Coachman *et al.*, 1975] Coachman and coworkers recognized the presence of two distinct water masses in the Chukchi Sea that might contribute to subsurface temperature maxima downstream in the Arctic Ocean. The first is Alaskan Coastal Water (ACW), a fresh ($31 < S < 32$) and relatively warm ($1^{\circ}\text{C} < T < 6^{\circ}\text{C}$) current containing significant river influence (especially, the Yukon River). ACW hugs the Alaskan coast at least until Point Barrow, where some fraction spins off into Canada Basin eddies [Belyakov and Volkov, 1982; D'Asaro, 1988], and the rest continues eastward along the continental slope [Aagaard, 1984], possibly shedding more eddies [Shimada, 2001]. The current appears far downstream at the continental slope north of Alert, Ellesmere Island [Newton and Sotirin, 1997], although it has lost much of its ACW temperature maximum by that point (see section 4.2). Its forcing mechanism is unclear, although an eddy-bathymetry interaction may be involved [Holloway, 1992]. Its barotropic component may be similar to the boundary current observed along the Eurasian continental slope [e.g., Woodgate *et al.*, 2001]. A baroclinic structure has been observed in the upper few hundred meters that Aagaard [1984] ascribed to geostrophic shear induced by pycnocline upwelling. This upwelling is forced by offshore Ekman transport in response to westward surface stress in the large-scale Beaufort Gyre. The resulting vertical shear across the halocline may mean that eastward transport of Pacific waters in this current is possibly quite sensitive to the large-scale surface stress forcing, i.e., to the strength of the Beaufort Gyre in a particular year.

[6] The second water mass discussed by *Coachman et al.* [1975] was referred to as Bering Sea Water (BSW), and occupies the bulk of the central Chukchi Sea. It is a merged product of Bering Shelf Water and Gulf of Anadyr Water, with a resulting summer salinity of $32 < S < 33$ and summer temperature of $0^{\circ}\text{C} < T < 2^{\circ}\text{C}$. This water mass drains northward, primarily through Herald Canyon and Hanna Canyon (also known as “Central Trough”), although some may also mix into the coastal current (R. Woodgate, University of Washington, and T. Weingartner, University of Alaska, personal communication, 2002). In this study, we adopt the nomenclature of *Coachman et al.* [1975]; that is, we retain the distinction between the fresher, generally warmer ACW and the saltier, often cooler BSW. We define sBSW as the summer variety of BSW, i.e., that which is heated and freshened in the Chukchi Sea during summer. Together, ACW and sBSW will be referred to as “summer Pacific halocline waters.” Starting in section 6, we further distinguish between summer BSW (sBSW) and winter BSW (wBSW).

[7] Recently, *Shimada et al.* [2001] examined data collected during the 1997–1998 drift of ice station SHEBA.

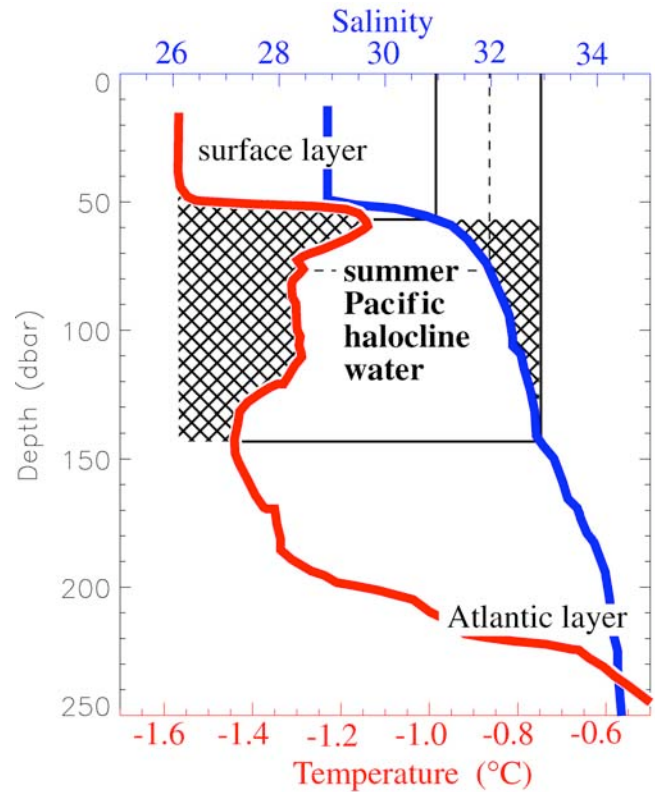


Figure 1. Temperature (T) and salinity (S) profiles from SCICEX'96, cast 43 (location shown in Figure 2). Following *Shimada et al.* [2001], summer Pacific halocline water is defined by a temperature maximum within the salinity range $31 < S < 33$. Hashed areas indicate the heat in the summer Pacific halocline water (relative to the surface layer) and the freshwater content (relative to $S = 33$).

Figure 3 shows an example from early April over the southern Chukchi Borderland, close in space and time to their “Station C”. (The Chukchi Borderland is here defined as the Northwind Ridge, the Chukchi Cap, and the intervening Northwind Abyssal Plain.) There are two clearly distinct temperature maxima, one each within the salinity ranges for ACW and sBSW. (*Shimada et al.* also observed a third, fresher temperature maximum formed by solar heating of a surface mixed layer heavily influenced by the Mackenzie River. This will not be discussed further in the present study.) These waters are influenced by air-sea exchange and other processes in the Chukchi Sea, and thus exhibit significant variability in space and time. Nonetheless, the temperature maxima were generally confined to the ACW and sBSW salinity ranges shown in Figure 3, at least for the SHEBA data. *Shimada et al.* [2001] also found that east of the Chukchi Borderland, the temperature maximum was predominantly ACW, which makes sense as it is closer to (although up to hundreds of kilometers away from) the Alaskan coast. Similarly, over the western Chukchi Borderland, sBSW dominated the temperature maximum. Thus, over this relatively limited region, three regimes of summer Pacific halocline waters were observed: Separated ACW and sBSW regimes in the east and west, respectively, and a combined ACW/sBSW regime in the center.

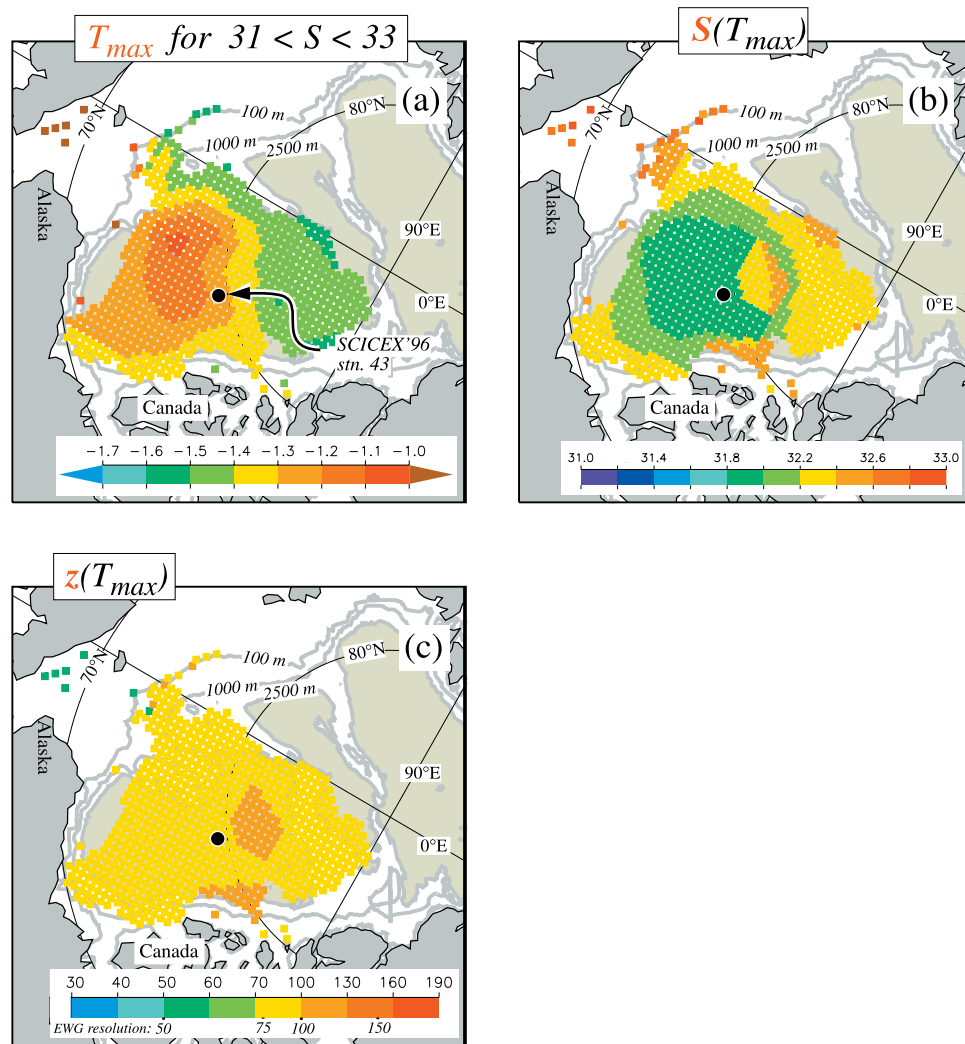


Figure 2. Lateral extent in the Arctic Ocean of (a) the summer Pacific halocline water temperature maximum ($^{\circ}\text{C}$), (b) the salinity of this maximum, and (c) the depth of this maximum (m), all computed using the 50-km gridded summer Arctic Ocean Atlas [EWG, 1998]. Grid points where the temperature maximum is colder than -1.6°C or shallower than 30 m are not plotted. The location of the profiles in Figure 1 is shown as a black dot. Also shown in Figure 2c is the vertical depth resolution of the EWG data set, which like most climatologies, poorly resolves the halocline. On this and subsequent maps, bathymetric contours are provided at 100, 1000, and 2500 m, with depths greater than 2500 m shaded.

[8] Is it possible to trace the circulation of these summer Pacific halocline waters downstream into the Arctic Ocean? We will examine this issue using a variety of hydrographic data sets collected mostly from the late 1980s to the present. Our tracer for summer Pacific halocline waters will be the presence of a temperature maximum within the salinity ranges defined by Shimada *et al.* [2001]. Analysis of other tracers such as dissolved nutrients has been successfully used to trace Pacific water influence in the upper Arctic Ocean [e.g., Jones *et al.*, 1998], and might prove useful in distinguishing Pacific halocline varieties, but is beyond the scope of this work. We will show how the circulation of summer Pacific origin waters responds dramatically to changes in wind patterns associated with large-scale climate modes. We then speculate how this variability might affect the outflow of these waters to the North Atlantic Ocean through the various straits of the Canadian Archipelago and Fram Strait.

[9] The following section describes the data and methods used in this study. We next analyze a variety of hydrographic and sea ice data from the Arctic Ocean, with a focus on changes observed north of Ellesmere Island, a key location for diagnosing large-scale circulation changes. These changes are then discussed in the context of climate mode variability. This leads us to reconsider the concept of “typical” Canadian basin temperature profiles, separated now into three regimes and including both summer and winter Pacific waters. We end with our conclusions, including a brief discussion of the variability of Pacific source waters.

2. Data and Methods

2.1. Data

[10] We have analyzed temperature and salinity profiles from a variety of sources over the years 1989–2001 (Figure 4).

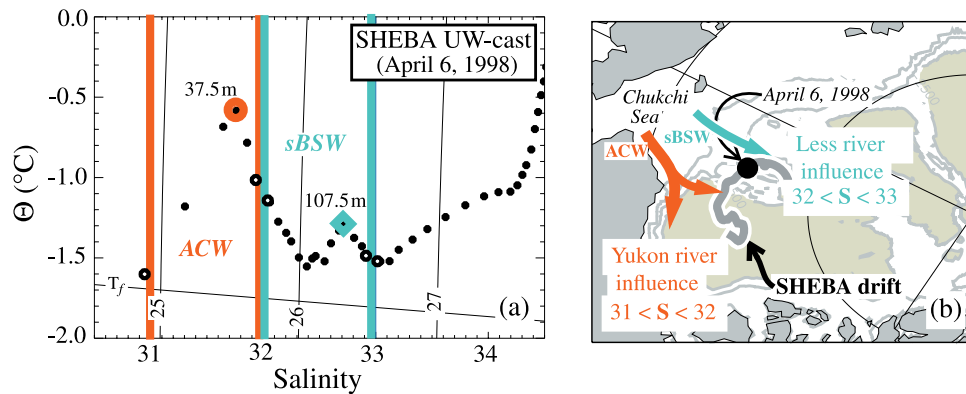


Figure 3. (a) θ -S plot from a SHEBA CTD cast taken by the University of Washington (UW) on 6 April 1998 in the south-central Chukchi Borderland, binned to 5-m resolution. The red dot and blue diamond show temperature maxima within the salinity ranges for Alaskan Coastal Water (ACW) and summer Bering Sea Water (sBSW), respectively. (b) The geographic position of this SHEBA station samples both ACW and sBSW outflows from the Chukchi Sea.

Coverage of the deep Arctic Ocean over the entire period is reasonably good, although individual years of course exhibit large gaps. Note the significant “data hole” north of the Canadian Archipelago and Greenland. We have compared these profile data with the gridded climatology known as the Arctic Ocean Atlas [EWG, 1998]. Also, we present an analysis of drifting sea ice buoy data, in order to provide a sense of upper ocean forcing and circulation.

[11] The oldest hydrographic data that we analyzed come from two remarkable time series of late winter conditions north of the North American coast. The first was obtained by Melling [1998] in the Mackenzie River delta and beyond over the years 1979–1996 (although here we present data only from 1993 and 1996). The second was collected on the Ellesmere Island shelf and continental slope by Newton and Sotirin [1997] as part of the ICESHELF project over the years 1989–1996. ICESHELF data from 1991 to 1996 were kindly provided to us by J. Newton (personal communication); data from 1989 to 1990 were estimated from figures by Newton and Sotirin [1997].

[12] We also analyzed profiles obtained by icebreaker cruises, including the Larsen’93 cruise [Carmack et al., 1995; McLaughlin et al., 1996], the JOIS’97 (leg 4) cruise on the CCGS *Louis S. St-Laurent* [Macdonald et al., 1999], the 1996 cruise of the USCGC *Polar Star* [Weingartner et al., 1998], and the R/V *Polarstern* cruises to the Eurasian Basin of the Arctic Ocean in 1993 (ARKIX-4) and 1996 (ARKXII) [Schauer et al., 2002]. In addition, profiles obtained by submarine in 1993, 1996, 1997, 1999, and 2000 as part of the U.S. SCICEX project were also analyzed [e.g., Morison et al., 1998; Steele and Boyd, 1998; Gunn and Muench, 2001; Smethie et al., 2000]. Data collected by the University of Washington during the 1997/1998 SHEBA drift of the CCGS *Des Groseilliers* were also used.

[13] The most recent data we analyzed come from hydrographic surveys conducted as part of the North Pole Environmental Observatory (NPEO), a 5-year (2000–2004) project sponsored by the National Science Foundation. NPEO is composed of several observational programs, all centered at the North Pole. The data discussed here were obtained using a Seabird SBE-19 Seacat conductivity-tem-

perature-depth (CTD) instrument that was lowered into the ocean through a hole in the sea ice. Transportation was by ski-equipped Twin Otter fixed wing aircraft based at Alert, Canada and the North Pole. Further details, including first

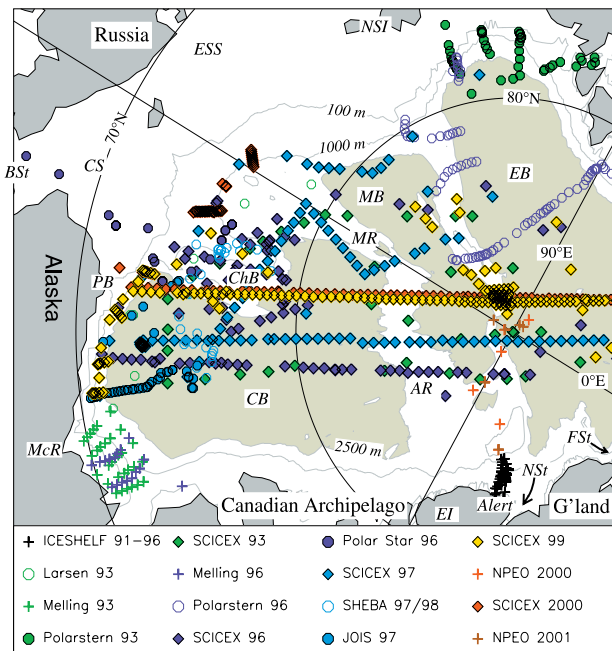


Figure 4. Hydrographic stations used in this study, color-coded by year. Icebreaker cruises are designated by circles (open or solid), submarine cruises by diamonds (open or solid), and aircraft-based operations by pluses. Geographic locations are marked as Bering Strait (BS), Fram Strait (FS), Nares Strait (NS), Chukchi Sea (CS), East Siberian Sea (ESS), Canada Basin (CB), Makarov Basin (MB), Eurasian Basin (EB), Mendeleyev Ridge (MR), Alpha Ridge (AR), New Siberian Islands (NSI), Point Barrow (PB), Mackenzie River (McR), and Ellesmere Island (EI). The Yukon River discharge enters the Bering Sea south of Bering Strait (not shown) and, under the influence of the Earth’s rotation, turns northward to flow into the Arctic Ocean through Bering Strait.

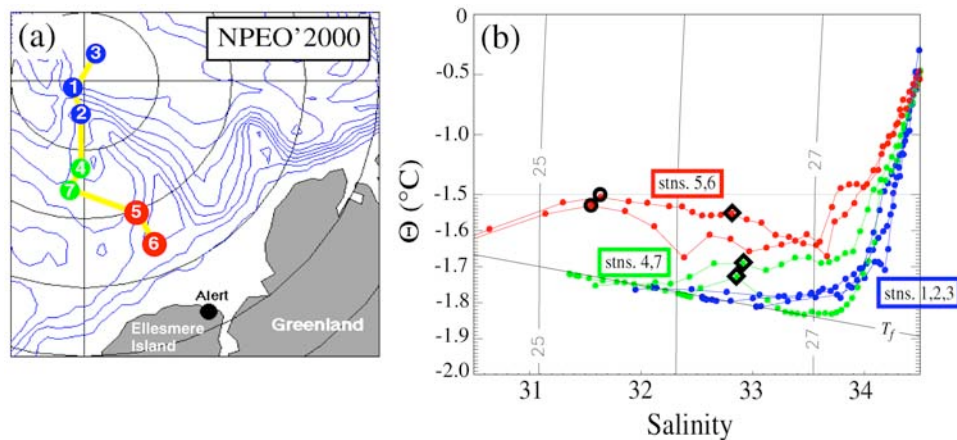


Figure 5. (a) NPEO hydrographic stations in the year 2000. (b) θ -S profiles (binned to 4 m but unsmoothed), grouped into three types, as discussed in the text. Note the break in temperature scale at -1.5°C . Temperature maxima for ACW (black circles) and for sBSW (black diamonds) are found at stations 4, 5, 6, and 7. No summer Pacific halocline water maxima are found at stations 1, 2, or 3. (See section 2.2 for a description of our temperature maximum detection algorithm.) The freezing line is marked as T_f .

scientific results, were published by *Morison et al.* [2002]; latest information is available at <http://psc.apl.washington.edu/northpole>.

[14] In 2000 (the first year of NPEO) a large-scale section of seven stations was made between Alert, Canada to the North Pole, and just beyond into the Amundsen Basin (Figure 5). These stations can be grouped into three main types, starting from the north. Stations 1, 2, and 3 are Atlantic water mass assembly (using the nomenclature of *McLaughlin et al.* [1996]), i.e., salty surface waters, no heat in the halocline, and warm Atlantic Water layer temperature maximum. As discussed by *Morison et al.* [2002], stations 4 and 7 are at the Atlantic/Pacific front; that is, they exhibit lower surface salinities, a very weak temperature maximum in the halocline, and significantly cooler Atlantic Water layer temperature maximum. Finally, stations 5 and 6 represent a Pacific water mass assembly, i.e., quite fresh surface salinity, a clear temperature maximum in the halocline, and a cool Atlantic Water layer. Station 6 (and NPEO'01 station 4 at the same location; see Figure 4) in particular represents a valuable data point, as it extends the annual survey taken in this general area each spring from 1989 to 1996 by *Newton and Sotirin* [1997].

[15] There are two distinct temperature maxima at NPEO'00 station 6, although they have nearly identical values. The result is a remarkably uniform and broad maximum over a salinity range of ~ 2 . This observation leads us to speculate that the broad temperature maximum has formed from two distinct water masses, i.e., from ACW and sBSW temperature maxima that have cooled during their transit from the Chukchi Sea to the area north of Ellesmere Island. This idea will be explored further below.

2.2. Methods

[16] Several methods might be applied to hydrographic profiles to distinguish ACW and sBSW. Here we have chosen to search for a potential temperature (θ) maximum within a specified salinity range in θ -S space. This is the core

property that we are seeking and is thus more specific than, for example, a mean θ over a salinity range. The presence of ACW is defined by a θ maximum within $31 < S < 32$, and sBSW by a θ maximum within $32 < S < 33$. The maxima were identified using an automated algorithm with no subjective user modifications. This has obvious advantages, but unfortunately does produce occasional false positive and false negative identifications. However, these generally do not affect the overall pattern of Pacific halocline water circulation. Results were similar when we used an alternate diagnostic, i.e., the mean temperature within ACW and sBSW salinity ranges. (In this paper “temperature” refers to the potential temperature relative to the surface. The difference between in situ and potential temperature is less than 0.01°C for the temperatures, salinities, and depths above 200 m that we consider here.)

[17] Maxima at the boundaries of a salinity range were allowed only if a lower temperature was found just outside the range, in order to eliminate false positives from monotonically increasing or decreasing θ -S curves. The data were first binned or linearly interpolated to 5-m resolution, then smoothed (with respect to salinity) with a three-point running median smoother. Further smoothing was not applied because it frequently eliminated a maximum, owing to the relatively few depth values that sometimes span the ACW and/or sBSW salinity ranges. The algorithm searches for the maximum value in these smoothed data, and returns the corresponding warmest unsmoothed data value. An example is provided in Figure 6. Our algorithm also identifies the depth and salinity of each temperature maximum. An example was provided in Figure 2 for climatological data.

3. Hydrographic and Sea Ice Data Analysis

3.1. 1993 Hydrographic Data

[18] Hydrographic data from the year 1993 provide unusually good spatial coverage of Pacific halocline water circulation in the central Arctic Ocean. Cruise data are

available from SCICEX'93, *Larsen*'93, ARKIX-4 (*Polarstern*'93), and the *Newton and Sotirin* [1997] and *Melling* [1998] time series. We use these data as a starting point to discuss summer Pacific halocline water circulation, and in subsequent sections also consider data from other years. We first consider the entire range of summer Pacific halocline water ($31 < S < 33$), and then partition the data into the separate salinity ranges for sBSW and ACW.

3.1.1. 1993 Summer Pacific Water Circulation

[19] Figure 7 shows the extent and geographic variations of summer Pacific halocline water over the entire range $31 < S < 33$. As did *Coachman and Barnes* [1961], we find the warmest Pacific waters closest to their source, with maximum values $\sim 1^\circ\text{C}$ above freezing near the Chukchi Borderland. A hint of two pathways for this water is evident, i.e., one pathway closer to the northern Alaska shelf, and the other straight over the Chukchi Borderland. This is similar to the two pathways proposed for Pacific water in the surface mixed layer by *Jones et al.* [1998], a point we will return to below.

[20] The Pacific water cools to $\sim 0.5^\circ\text{C}$ above freezing in much of the Canada Basin. A front at about -1.4°C is clearly evident, aligned at a shallow angle to the Alpha-Mendeleyev Ridge. Profiles on the other side of this front show either a much cooler maximum, or no maximum at all. Note that this is not quite the pattern for cooling proposed by *Coachman and Barnes* [1961], who instead postulated that the cooling should progress along an anticyclonic pathway, following the Beaufort Gyre.

[21] Figure 7 also shows the depth and salinity of the temperature maximum. In much of the Canada Basin, the maximum lies between 50–70-m depth and 31.4–32. There is a hint of a deepening of this maximum in the middle of the Beaufort Gyre (at SCICEX'93 station 30, yellow symbol in Figure 7c), possibly owing to Ekman convergence within the anticyclonic Beaufort Gyre. Also, values near the outer edge of Pacific influence are generally deeper (70–100 m) and saltier (32–33). A comparison with Figure 2 indicates that the 1993 temperature maximum was generally shallower than indicated by the climatology, possibly the result of low vertical resolution in the latter.

[22] What might cause the spatial variations evident in Figure 7? Our leading hypothesis is that the summer Pacific halocline water temperature maximum acts as a tracer for the circulation of this water mass, decreasing in amplitude with increasing distance away from its source in the northern Chukchi Sea. We will assume a fixed input at this source region, and ascribe interannual variations within the Arctic Ocean to changes in the local circulation. As we shall show, this produces a consistent picture of the circulation. In the Conclusions, we briefly discuss alternate hypotheses.

3.1.2. Partition Into sBSW and ACW

[23] *Shimada et al.* [2001] showed that summer Pacific halocline waters can be partitioned into two subspecies, which we are referring to here as ACW and sBSW. Figure 8 demonstrates the application of this concept to the 1993 cruise data, indicating that the two pathways hypothesized by *Jones et al.* [1998] are in fact composed of different source waters. The branch that lies closer to the coast is ACW, which dominates the data in the southern Beaufort Sea. The branch that transits directly over the Chukchi Borderland is mostly sBSW, which extends farther away from the Chukchi

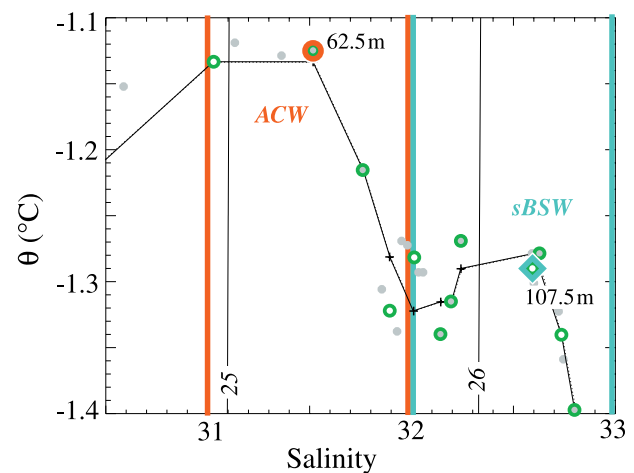


Figure 6. An example of our temperature maximum algorithm, applied to the profile shown in Figure 1. The raw data (gray dots) are first averaged into 5-m bins (green circles), then smoothed with respect to salinity using a three-point running median smoother (black pluses, connected by the black line). If a temperature maximum is found in this smoothed curve, the corresponding unsmoothed binned value is designated as a temperature maximum. In this example, both ACW (red circle) and sBSW (blue diamond) maxima were found.

Sea source region toward the Eurasian Arctic. A broad region of overlap in the Canada Basin exists in which both ACW and sBSW are present. Throughout much of the Canada Basin, ACW generally resides at 40–70-m depth, while sBSW lies just below at 70–130 m. There is some indication of a shoaling of sBSW toward the Makarov Basin, possibly where it rises over halocline water of Atlantic origin.

[24] Figure 8 provides an explanation for the spatial variations of Figure 7, specifically for the changes in summer Pacific water properties at the outer (Eurasian) boundary of its domain. If one searches for a single temperature maximum over the entire range $31 < S < 33$, then the search will find only ACW where the cooler sBSW and the warmer ACW overlap. Thus sBSW is hidden from view until the ACW influence ebbs at its outer boundary near the Makarov Basin. At this point sBSW becomes visible, creating the sudden transition to deeper, saltier values in Figure 7.

[25] The *Melling* [1998] data from the Mackenzie River delta show a clear ACW influence and an indeterminate sBSW influence. This probably indicates the presence of an eastward flowing boundary undercurrent along the North American continental slope, such as first proposed by *Aagaard* [1984]. However, the data also show a clear ACW presence throughout the deep Beaufort Sea, indicating that a pathway to deep water also exists. The *Newton and Sotirin* [1997] data on the Ellesmere Island continental slope show the presence of both types of summer Pacific halocline water, although the ACW is quite weak. The sBSW temperature maximum is much warmer than observed in 2000 during NPEO (Figure 5), the cause of which will be explored in section 4.

3.1.3. Circulation Schematic for the Year 1993

[26] Figure 9 presents a schematic flow diagram for summer Pacific halocline water circulation in the year

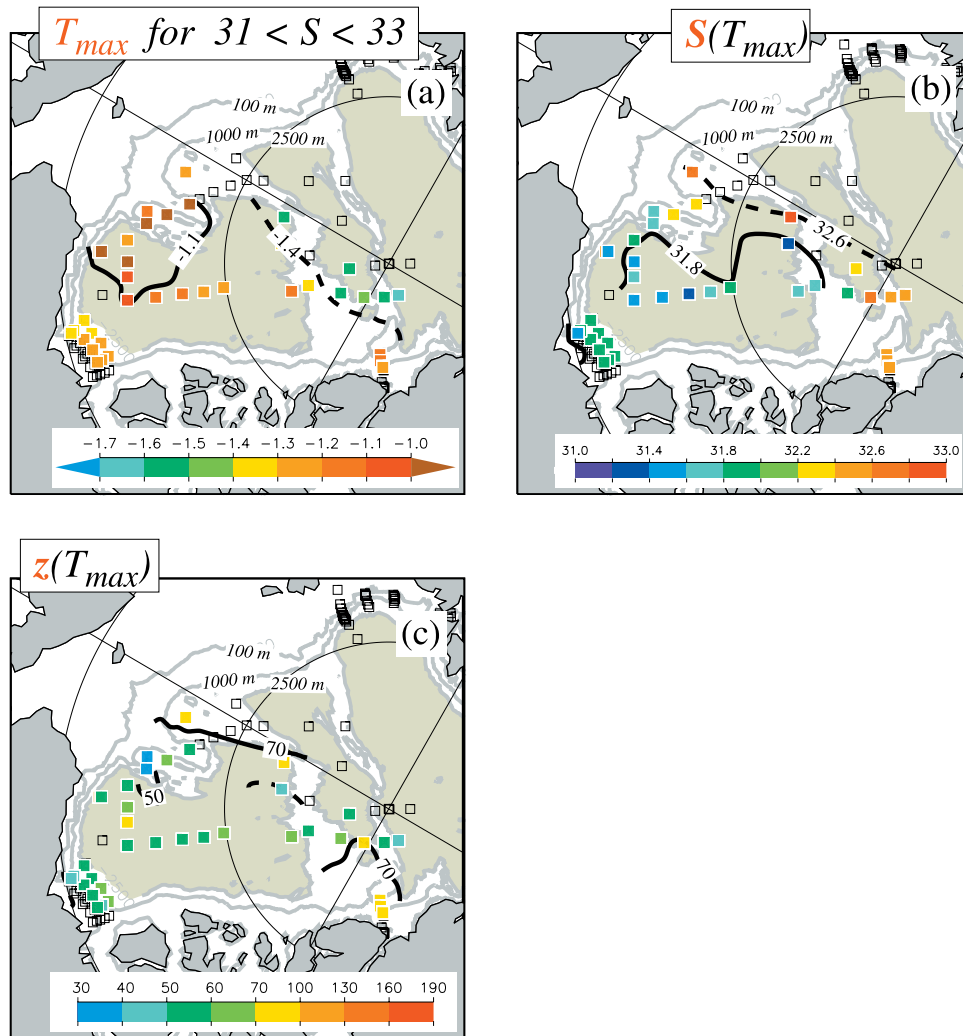


Figure 7. (a) Temperature maximum ($^{\circ}\text{C}$) over the entire summer Pacific halocline water salinity range $31 < S < 33$ for 1993 cruise data from SCICEX'93, Larsen'93, and ARKIX/4 and time series from Newton and Sotirin [1997] and Melling [1998]. Stations where there was no salinity in this range are not plotted. Also shown are (b) the salinity and (c) the depth (m) of the temperature maximum. Open symbols denote profiles where there is no maximum in this salinity range or where the maximum is in the mixed layer (i.e., both colder than -1.6°C and shallower than 30 m) or where the depth of the water is less than 500 m.

1993. This water is partitioned into sBSW and ACW flows. The figure presents four regimes, each illustrated with a temperature profile. The first regime is in the Eurasian sector of the Arctic Ocean, where no summer Pacific water is observed. The second regime lies near the outer boundary of summer Pacific water influence, where the Transpolar Drift Stream carries only sBSW directly from the Chukchi Sea. The third regime occupies much of the Canada Basin, where the warmer, fresher ACW lies just below the surface mixed layer and just above the cooler, saltier sBSW. The fourth regime lies in the southern Beaufort Sea, where ACW circulates in the Beaufort Gyre and within the boundary undercurrent.

[27] Details about how sBSW and ACW are injected from the Chukchi Sea into the Arctic Ocean are beyond the scope of this work. However, Figure 9 presents a speculative picture of some aspects of this problem. These summer

waters are highly stratified, and yet bathymetric steering is generally quite strong in the highly barotropic arctic seas. Thus we have drawn sBSW entering the Arctic Ocean through Herald Canyon and Central (also known as Hanna) Canyon, and ACW through Barrow Canyon. The real flow is probably not so simple. In fact, we expect that stratification in the Chukchi Sea decouples the surface flows from the bathymetry to some extent, possibly allowing significant mixing between these three idealized pathways. One observation that may support this idea is the weak sBSW signal in the Mackenzie delta (Figure 8), which might indicate a mixing of this water into the boundary undercurrent.

[28] ACW injection into the Beaufort Gyre is probably accomplished mostly by eddy transport from Barrow Canyon [D'Asaro, 1988], and from the boundary undercurrent north of Alaska [Shimada, 2001]. This is illustrated in Figure 9. Eddies may also play a role in the transport of

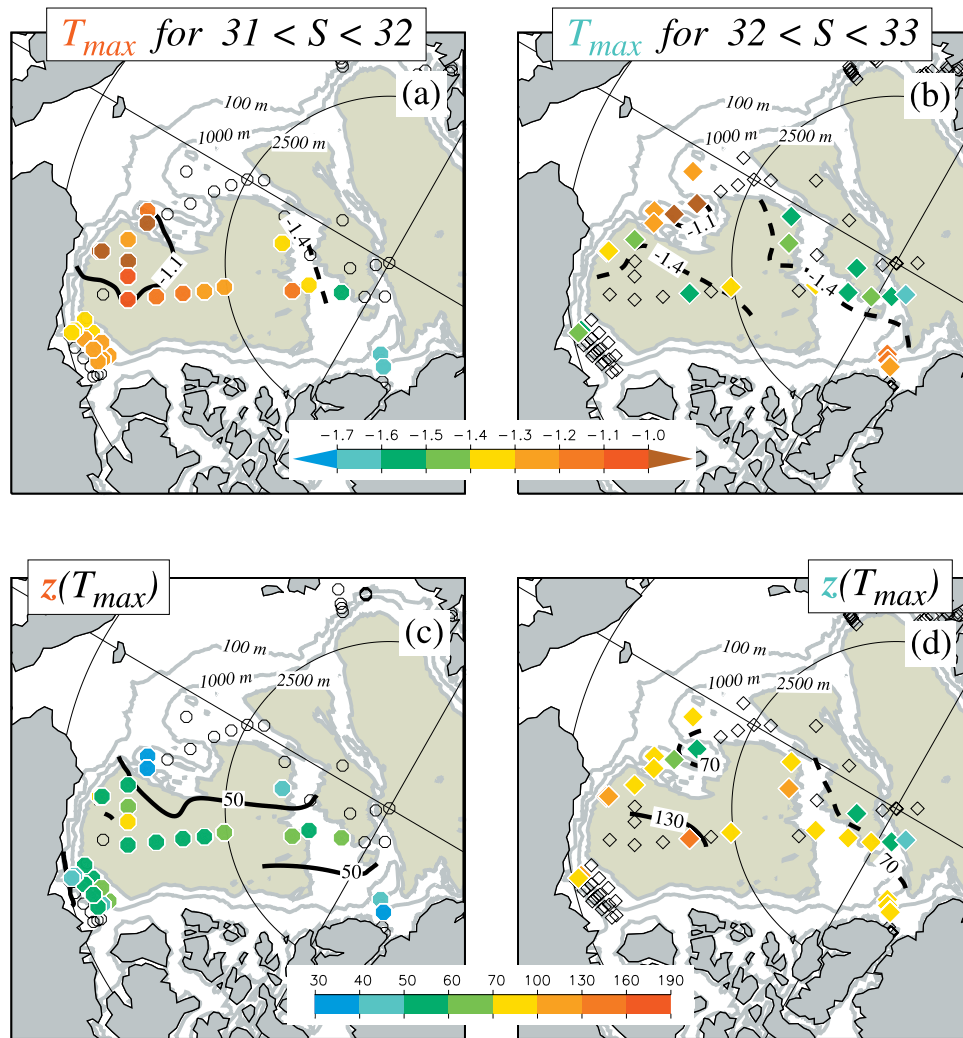


Figure 8. As in Figure 7 but with the two temperature maxima separated into the salinity ranges for ACW ($31 < S < 32$) and for sBSW ($32 < S < 33$): (a) and (b) the temperature maxima, T_{max} (°C) and (c) and (d) the depth of the temperature maxima, $z(T_{max})$ (m).

sBSW off the Chukchi shelf, an issue that has not been explored to date.

3.2. Post-1993 Hydrographic Data

[29] In this section we extend our analysis of hydrographic data into the latter half of the 1990s, in order to determine the interannual variability of summer Pacific water circulation. Figure 10 shows sBSW and ACW temperature maxima for the years 1996/1997 and 1999/2000. We have used 2-year pairs to increase data coverage. Coverage in 1996/1997 is reasonably good. The year 1996 was the last year for both the *Melling* [1998] and *Newton and Sotirin* [1997] time series. In general, the 1996/1997 data have more spatial variance than the 1993 data. This may result from the transition to different climate states during this time (see next section), or from eddies and other small-scale phenomenon that were not resolved by the coarser-scale 1993 data.

[30] Coverage in 1999/2000 is not nearly as good as in 1993 or 1996/1997. One reason is that the most common cross-arctic section taken by the SCICEX program was on

the Russian side of the North Pole, as shown in 1999/2000. This provides poor coverage of the deep Canada Basin. Another reason for reduced coverage is the end of the *Melling* [1998] time series north of the Mackenzie delta. On the other hand, after a 4-year break, the *Newton and Sotirin* [1997] time series was fortunately extended by NPEO, starting in 2000.

[31] Figure 10 shows that the basic distribution of ACW and sBSW does not change in the later 1990s, i.e., the ACW maximum lies closer to the Alaskan coast, while the sBSW maximum lies in the middle of the Canadian Basin. However, differences are also apparent. One example is a striking change that takes place in the Chukchi Abyssal Plain, east of the Chukchi Borderland. In 1993 (Figure 8) this area has a few sBSW observations and no ACW observations at all. In later years (Figure 10) a progressively warmer signal from both water masses appears, especially in ACW. Another area with significant change is north of Ellesmere Island, where throughout the later 1990s, the sBSW signal weakens (substantially) while the ACW signal strengthens (slightly). This results in a weak signal

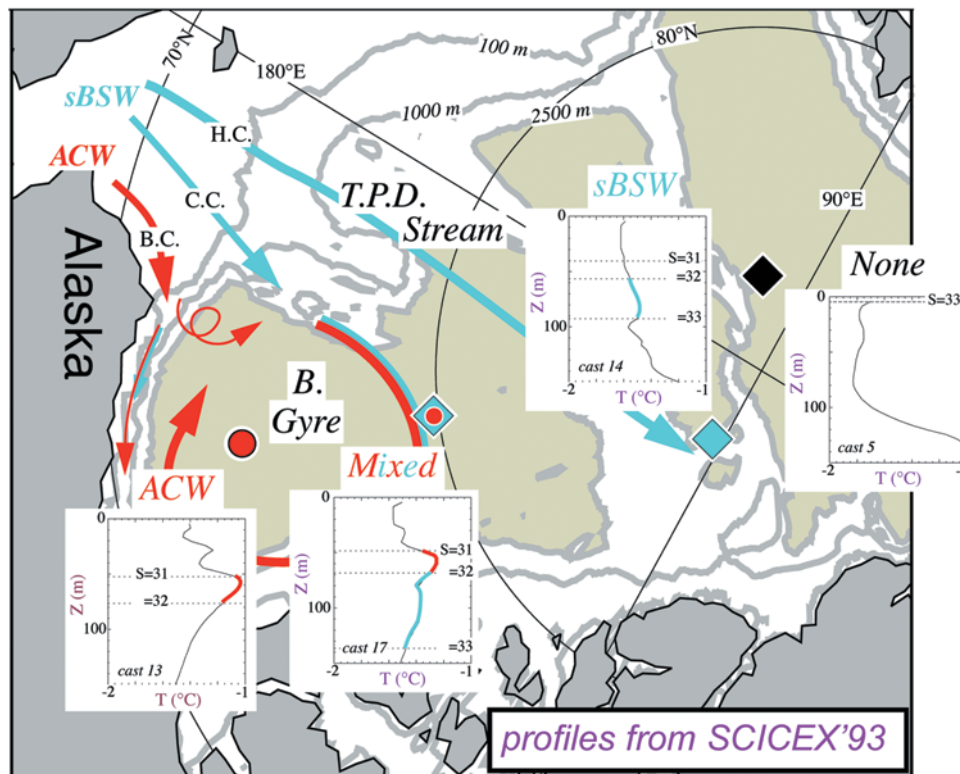


Figure 9. Schematic circulation of summer Pacific halocline water in the Arctic Ocean during the year 1993. Summer Bering Sea Water (sBSW, blue) enters the Arctic Ocean via Herald Canyon (H.C.) and Central Canyon (C.C.) and is swept up into the Transpolar Drift Stream (T.P.D. Stream) and the northern edge of the Beaufort Gyre (B. Gyre). Arctic Coastal Water (ACW, red) enters the Arctic Ocean via Barrow Canyon (B.C.) and continues along the shelf break in the coastal undercurrent. (Some BSW might also enter via this pathway.) ACW also spins northward off the coast in eddies (curly line) and enters the Beaufort Gyre. Example temperature profiles are shown from SCICEX'93 for each of four regimes, i.e., only ACW (red dot), only sBSW (blue diamond), mixed ACW and sBSW (red dot in blue diamond), and no summer Pacific halocline water temperature maximum (black diamond).

from both water masses at NPEO'00 station 6, as noted in section 2.

[32] Figure 10 also shows good cross sections of the boundary current north of Alaska, from the Canadian JOIS'97 cruise and to a lesser extent, the SCICEX'99 cruise. There is an ACW temperature maximum near the shelf break in 1997 (Figures 10a and 10b), with lower values seaward of this position, and then a strengthening again within the Beaufort Gyre. In 1999 (Figures 10c and 10d) the data suggest ACW penetration along the Alaskan continental slope, and also a strengthening sBSW influence. Still, sBSW is cooler than ACW within this current, and only reaches maximum temperatures within the central Canadian Basin. This suggests that its main pathway may be over the Chukchi Borderland and into the deep Canada Basin.

3.3. Sea Ice Circulation and the Arctic Oscillation

[33] Our analysis of hydrographic data from the 1990s reveals a changing distribution of summer Pacific halocline water influence. Might these changes be forced by climate mode fluctuations? In particular, we seek a correlation with the Arctic Oscillation (AO), which is the first component of an Empirical Orthogonal Function of northern hemisphere sea level pressure, explaining 52% of the variance over the

wintertime Arctic Ocean [Rigor *et al.*, 2002]. Our hypothesis is that surface pressure anomalies will cause surface wind anomalies, which in turn will force sea ice and upper ocean circulation anomalies that will impact Pacific water circulation patterns.

[34] Figure 11a shows AO monthly and mean January–February–March (JFM) time series. Also shown is the 3-year running mean of the JFM time series. “Positive”, “negative”, and “neutral” AO indices are all taken here relative to the 1979–2001 mean. The well-known sharp increase in AO index in the late 1980s is evident. Also evident is a gradual decline toward lower index values in the later 1990s. One may alternatively create AO time series using other groupings of months, for example, November–May or January–December annual means. These yield generally the same qualitative behavior as in Figure 11a, although there are differences. For example, the annual mean AO index time series (not shown) is very similar to the JFM time series in the 1980s, but is generally lower in value in the 1990s, especially the later 1990s. This may result from the influence of other climate modes during this time, a topic that is beyond the scope of this study.

[35] We have color-coded three periods in Figure 11a: 1979–1987 (green) has a generally negative index, 1988–

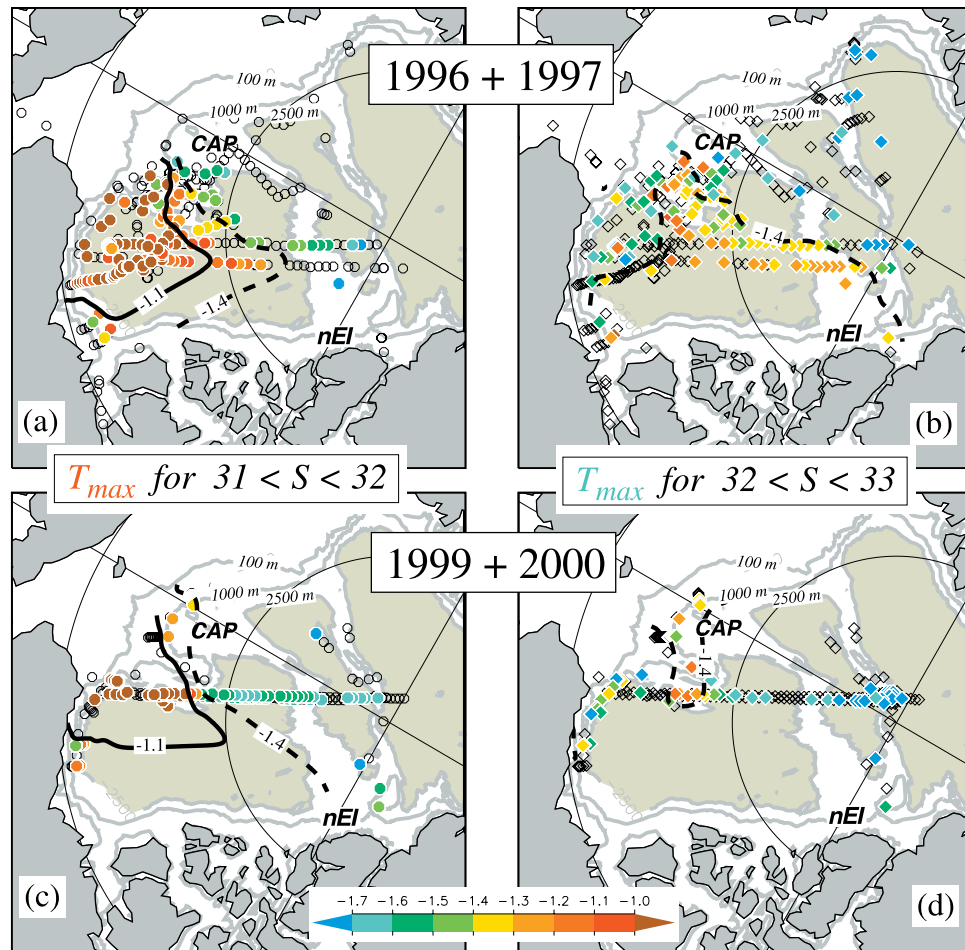


Figure 10. As in Figure 8 but only for $T_{\max}$ ($^{\circ}\text{C}$) and for the years (a) and (b) 1996/1997 and (c) and (d) 1999/2000. Data are from submarine cruises SCICEX'96, '97, '99, and '00, from icebreakers *Polarstern*'96 (ARKXII), *Polar Star*'96, and JOIS'97, and from aerial surveys/camps ICESHELF'96, Melling'96, and NPEO'00 and '01. Areas with distinct change over the years 1993–2000 (discussed in section 6) are marked on the figure as the Chukchi Abyssal Plain (CAP) and the area north of Ellesmere Island (nEI).

1994 (red) has a strongly positive index, and 1995–2001 (blue) has a weakly positive index. In Figures 11b, 11c, and 11d, we have then formed maps of annual (not just JFM) mean sea ice motion and sea level pressure over these three time periods, using data from the International Arctic Buoy Program [e.g., *Rigor et al.*, 2002]. We have also “seeded” these fields with Lagrangian tracers that originate every month on an Eulerian (fixed) grid and are advected with the mean sea ice motion over each of the three time periods. Following the arguments by *Morison et al.* [1998], we make the simple assumption that this provides at least a first-order tracer for the upper ocean circulation. Further work using numerical ice-ocean models would be useful in this regard.

[36] The main differences in these three time periods appear in the size and position of the Beaufort Gyre and the Transpolar Drift Stream. In the earliest, negative AO period (1979–1987, Figure 11b) the Transpolar Drift Stream originates around the New Siberian Islands. The Beaufort Gyre is quite large and strong, which leads to strong westward ice motion north of the Chukchi Sea and the Alaskan coast, where Pacific waters enter the Arctic

Ocean. Significant westward motion is also evident along the Canadian continental slope, all the way to Ellesmere Island. The next, strongly positive AO period (1988–1994, Figure 11c) is very different. The origin of the Transpolar Drift Stream is shifted further east toward the East Siberian Sea. The Beaufort Gyre is very much weaker, which means weaker westward winds north of the Chukchi Sea and the North American continental slope. In fact, eastward ice motion is evident along the eastern Canadian Arctic Archipelago, which is part of the Transpolar Drift Stream that has shifted toward North America [e.g., *Steele and Boyd*, 1998]. The final, weakly positive AO period (1995–2001, Figure 11d) has properties that lie in between the two extremes. The Transpolar Drift Stream origin is rotated back toward the New Siberian Islands, but not as far as in the first period (Figure 11b). The Beaufort Gyre is strengthened, with increasing westward ice motion in the northern Chukchi Sea and the North American continental slope. However, ice velocities are not as high as in the first period. The motion north of Ellesmere Island is very weak, with little east-west component at all.

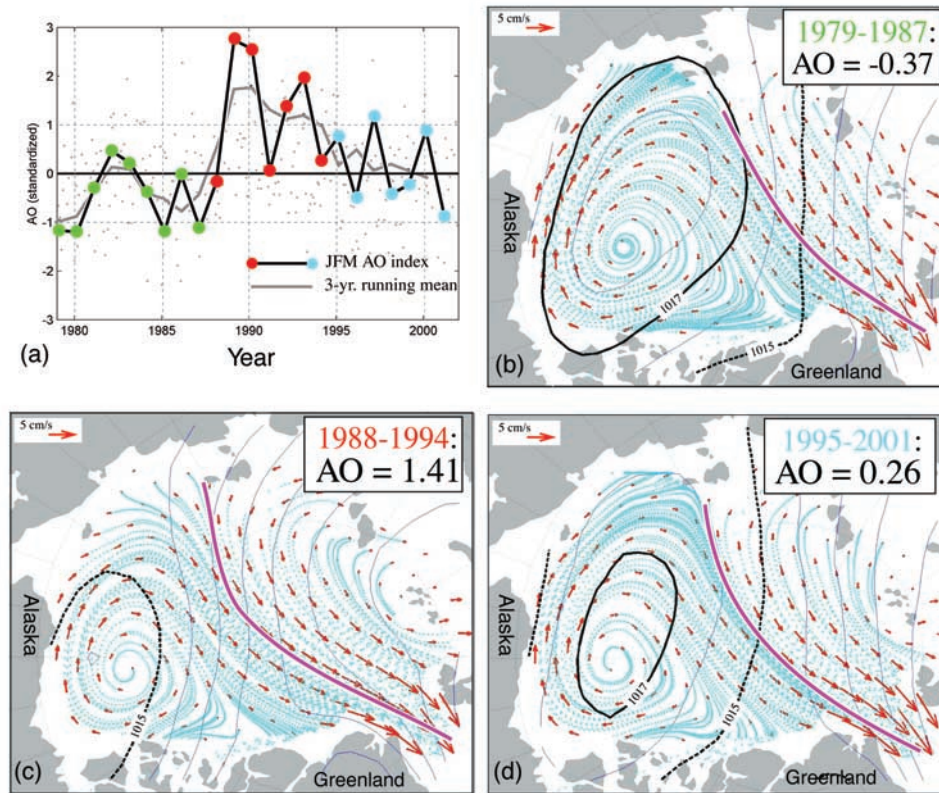


Figure 11. (a) Time series of the Arctic Oscillation (AO) index for the months January–March (JFM) and its 3-year running mean. Also shown are the monthly values (small gray dots). The JFM values have been color-coded into three periods: 1979–1987 (green), 1988–1994 (red), and 1995–2001 (blue). (b)–(d) The mean sea ice motion (vectors) and sea level pressure (contours) in these three time periods (using all months). Pressure contours at 1015 (dashed) and 1017 mb (solid) are marked to highlight the changes over these years. Blue streaks are Lagrangian tracer pathways forced continuously by these mean vector fields. The zero vorticity contour (magenta) indicates the main axis of the Transpolar Drift Stream. Also shown are the values of the mean JFM indices over these three time periods. Since 1979 the Arctic Ocean has experienced a period of negative AO index, then strongly positive AO index, and more recently, a weakly positive AO index.

[37] Buoy data also show (Figures 12a, 12b, and 12c) that the sea ice originating in the northern Chukchi Sea (74°N , 165°W) takes one of several different pathways, depending on climate forcing. In low AO index years, it either enters the Russian side of the enlarged Beaufort Gyre and takes ~ 6 years to cross the Arctic Ocean (Figure 12a), or it moves quickly into the East Siberian Sea and melts (not shown). In high AO states, it either takes a quick ~ 3 years to move across the Arctic Ocean in the Transpolar Drift Stream (Figure 12b, blue curve) or it becomes entrained into the reduced Beaufort Gyre and completes at least one full rotation around the gyre (Figure 12b, green curve). In neutral AO states, the ice may follow any of these pathways; when it passes near Ellesmere Island, it has generally taken ~ 4 years to get there (Figure 12c).

4. Changes North of Ellesmere Island

[38] The recent changes in ice and upper ocean circulation that we have shown should be most evident in certain key locations. In section 3.2, we discussed changes in the Chukchi Abyssal Plain and the area north of Ellesmere

Island. In this section, we focus on the latter region, where we have a relatively long hydrographic time series.

4.1. Sea Ice Trajectories

[39] The area north of Ellesmere Island is influenced by ice and water that originates in very different locations, depending on the climate state. This is illustrated in Figures 12d, 12e, and 12f, which show backward sea ice trajectories from the point 85°N , 65°W , using buoy data from the same three time periods as in Figure 11. Figure 12d confirms that this area is influenced by sea ice from mid-Siberia during negative AO index years (1981–1987). On the other hand, Figure 12e shows that during high AO index years (1988–1994), the sea ice in this area originates in the Beaufort Gyre and passes over the northern Chukchi Sea shelf, where the underlying ocean currents likely pick up a strong Pacific signature. Figure 12f shows anticyclonic motion as the ice moves in the Beaufort Gyre during neutral AO index years (1995–2001), but with perhaps less direct Chukchi influence.

[40] The buoy data thus indicate that the area north of Ellesmere Island may have had very little Pacific influ-

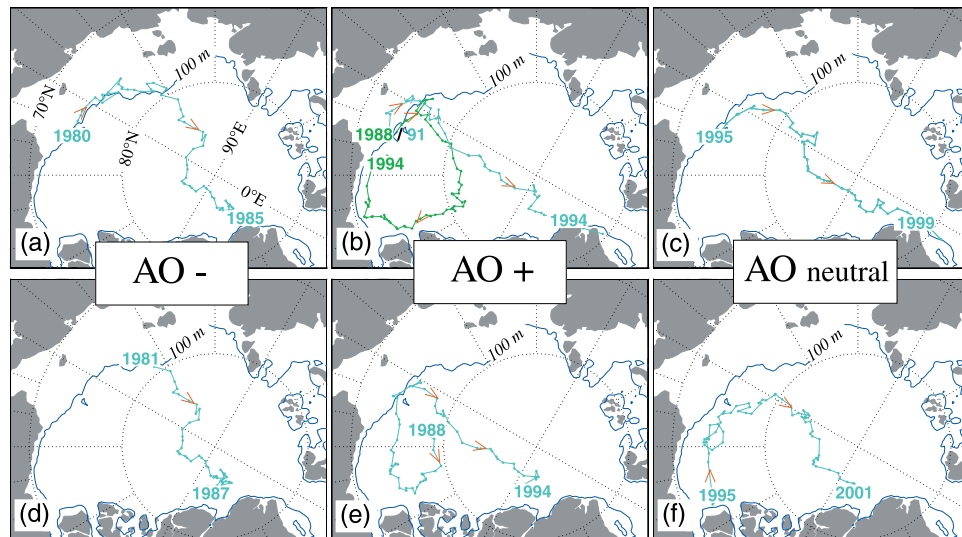


Figure 12. Sample sea ice trajectories (using buoy data) during negative, positive, and neutral phases of the Arctic Oscillation (i.e., the same three time periods used in Figure 11). (top) Trajectories that originate at the northern Chukchi Sea shelf break (74°N , 165°W) in January of (a) 1980, (b) 1988 and 1991, and (c) 1995. The 100-m bathymetry contour is shown to illustrate the shelf break. (bottom) Trajectories that have been calculated backward from a location north of Ellesmere Island (85°N , 65°W), starting in April of (d) 1987, (e) 1994, and (f) 2001. All trajectories terminate after 6 years, or earlier if they exit the Arctic Ocean or encounter land.

ence during much of the 1980s (low AO index), or perhaps that the Pacific waters in its vicinity were relatively old and thus had lost their temperature maximum signal. When the AO index abruptly shifted to highly positive values in the late 1980s, we expect that a new Pacific water signature should appear within several years. Finally, in the latter half of the 1990s, we expect a relaxation back toward less Pacific influence in the Ellesmere region. In the following section, we use an oceanographic time series to confirm and expand on these ideas.

4.2. Hydrographic Time Series

[41] Figure 13 presents a time series of sBSW temperature maximum from the area north of Ellesmere Island, Canada. Here, we have combined springtime data taken over the continental slope (centered at 84°N , 65°W) from 1989–1996 taken by *Newton and Sotirin* [1997] with the southernmost station taken by NPEO [Morison *et al.*, 2002] in 2000 (station 6) and in 2001 (station 4; both NPEO stations were located at about 85°N , 67°W). Gaps in this time series occur in 1995 and 1997–1999. Only data from stations where the water depth was greater than 500 m (according to the IBCAO bathymetry, see <http://www.ngdc.noaa.gov/mgg/bathymetry/arctic/arctic.html>) were analyzed. The NPEO data lie ~ 100 km to the north of the *Newton and Sotirin* [1997] data, and thus probably do not directly sample the boundary undercurrent itself, but rather the general hydrographic conditions in this region.

[42] *Newton and Sotirin* [1997] noted an increase in summer Pacific halocline water signature in their data over the period 1989–1993, although no distinction between ACW and sBSW was made. Our examination

of their data reveals that (according to our temperature-salinity criterion) it is mostly sBSW. *Newton and Sotirin* [1997] also noted that in 1994 (the last year discussed in their study) the summer Pacific water temperature

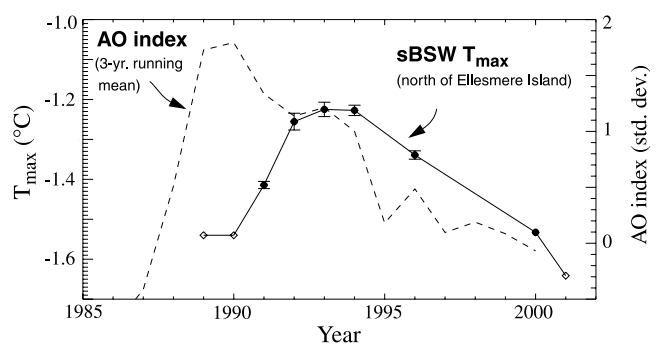


Figure 13. Time series of the mean sBSW temperature maximum (± 1 standard deviation) from north of Ellesmere Island, using *Newton and Sotirin* [1997] (1989–1996) and NPEO (2000–2001) observations. No standard deviation is plotted in 2000, when only one observation was taken. The open symbols in 1989–1990 designate years when no temperature maximum was observed, although an inflection point “plateau” exists [see *Newton and Sotirin*, 1997, Figure 9d], the temperature of which is plotted here. No temperature maximum or inflection point exists in 2001; in this year we have plotted the mean temperature within $32 < S < 33$. Also shown (dashed line) is the 3-year running mean of the Arctic Oscillation index (January–March mean), as in Figure 11a. The smoothed AO index has a maximum correlation of $r = 0.8$ with the Ellesmere Island data at a 3-year lag.

reversed its trend and actually decreased slightly. This decrease becomes more obvious in Figure 13, where we have combined *Newton and Sotirin* [1997] data with their (previously unpublished) data from 1996, and with NPEO data from 2000 and 2001. The beginning and end of this time series (i.e., years 1989, 1990, and 2001) show no evidence of any summer Pacific halocline water temperature maximum.

[43] The overall trend of the sBSW temperature maximum north of Ellesmere Island is a rapid increase in the early 1990s, followed by a gradual decline through the later years. This is clearly correlated with the well-known behavior of the AO, also plotted in Figure 13. We thus speculate that the rapid increase in the sBSW temperature maximum north of Ellesmere Island results from advection via the Transpolar Drift Stream, as its origin shifted from the New Siberian Islands during the negative AO index circulation regime of the 1980s (Figures 11b and 12d), to a point closer to the Chukchi Sea in the positive AO index regime of the late 1980s and early 1990s (Figures 11c and 12e). Similarly, the gradual decline of sBSW temperature maximum in this region would then be a result of the shifting back again of the Transpolar Drift Stream origin toward the New Siberian Islands and away from the Chukchi Sea source of this water in the later 1990s.

[44] A weak increase in the ACW temperature maximum during the early 1990s is also evident in the *Newton and Sotirin* [1997] data, starting from no signal in 1989 and 1990. It persists north of the continental slope in the NPEO data from 2000 (Figures 5 and 8). The origin of this signal is not clear. In positive AO years, its origin close to Alaska probably means that it is not entrained into the Transpolar Drift Stream, but rather into the Beaufort Gyre and the North American boundary current. The latter may (or may not) bring ACW all the way to the area north of Ellesmere Island. In negative AO years, on the other hand, the enlargement of the Beaufort Gyre might sweep some ACW into the Transpolar Drift Stream and thence to Ellesmere Island.

4.3. Heat Loss in Summer Pacific Halocline Waters

[45] How quickly do summer Pacific halocline waters lose their heat on the way to the area north of Ellesmere Island? First, consider the cooling of sBSW. We assume a typical sBSW temperature profile $T(z)$ with a temperature maximum of -1°C at about 100-m depth, and minima of -1.5°C at 15 m shallower and -1.7°C at 15 m deeper. This layer contains about 40 MJ m^{-2} of heat, relative to a profile with no temperature maximum between the upper and lower points. This may be compared with the 140 MJ m^{-2} of heat in the generally warmer ACW layer, as calculated by *Shimada et al.* [2001]. The loss of heat from the sBSW layer F_{sBSW} is given by

$$F_{sBSW} = \frac{\partial}{\partial t} (\rho c_p T dz) = \rho c_p k \frac{\partial T}{\partial z}, \quad (1)$$

where ρ and c_p are ocean density and heat capacity, respectively, and k is the vertical diffusivity. Our “typical” profile yields a sBSW temperature gradient of about $0.6^{\circ}\text{C}/15 \text{ m}$. *Dewey et al.* [1999] estimate the upper halocline vertical diffusivity $k = 10^{-6} - 10^{-5} \text{ m}^2 \text{ s}^{-1}$ in the

Eurasian Basin of the Arctic Ocean, where the stratification is weaker than in the Canadian Basin, and the heat source is warm Atlantic Water. *Macdonald and Carmack* [1993] estimate a higher bulk diffusivity in the Canada Basin halocline of about $4 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$. Combining these two estimates, equation (1) then yields $F_{sBSW} = 0.2 - 8 \text{ W m}^{-2}$, which would completely deplete the sBSW of its heat content in roughly 0.25–7 years.

[46] We can narrow these estimates further by an examination of the synoptic sBSW data. *Morison et al.* [1998] calculate geostrophic current velocities at sBSW depth ($\sim 100 \text{ m}$) of $\sim 1 \text{ cm s}^{-1}$ during the AO+ year of 1993. This implies a transit time of ~ 5 years for sBSW to arrive at the area north of Ellesmere Island via the $\sim 1500\text{-km}$ Transpolar Drift Stream pathway from the northern Chukchi Sea, i.e., a bit slower than the sea ice (Figures 11c and 12b). The hydrographic data show that the sBSW temperature maximum persists (although cooled down) north of Ellesmere Island during these years. This means that the time to deplete sBSW of its heat content is no less than ~ 5 years, which when combined with our estimate from equation (1) indicates a heat loss time of 5–7 years. The associated sBSW heat loss rate is then $F_{sBSW} \leq 0.25 \text{ W m}^{-2}$, and the vertical diffusivity $k \leq 1.5 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$. This is roughly consistent with the figure of 0.3 W m^{-2} calculated by *Morison and Smith* [1981] for the heat lost from summer Pacific halocline water in this area at the ice island T3.

[47] Figure 12a indicates that during AO– years, sea ice takes about 6 years to travel from the northern Chukchi Sea via the Transpolar Drift Stream to Ellesmere Island, versus the 3 years it takes during AO+ years (Figure 12b). The transit time for the sBSW layer also presumably slows down somewhat from the 5 years it takes during AO+ years, perhaps to $5 \cdot (6/3) = 10$ years. The hydrographic data indicate that this is long enough for the sBSW temperature maximum to disappear. Thus we see that the survival of a sBSW temperature maximum along the Transpolar Drift Stream pathway all the way to the area north of Ellesmere Island (and presumably downstream into Nares and Fram Straits) is highly sensitive to the climate state.

[48] What about the other pathway, i.e., the North American boundary current? Our limited data indicate that if any summer Pacific halocline water survives to the area north of Ellesmere Island along this pathway, it is ACW, not sBSW. ACW contains more heat than sBSW; however, it may also experience more rapid heat loss since it resides just under the surface mixed layer. Its speed (and indeed, direction) probably varies quite dramatically in response to the strength of the Beaufort Gyre (i.e., the surface stress forcing) which creates a surface westward baroclinic component in addition to the eastward barotropic component. Given these uncertainties, we hesitate to estimate the heat loss characteristics of this current, and look to future studies to examine this issue in more detail.

5. Summer Pacific Halocline Water Circulation and the Arctic Oscillation

[49] Figure 14 shows our attempt to synthesize the observations into a schematic circulation scheme for

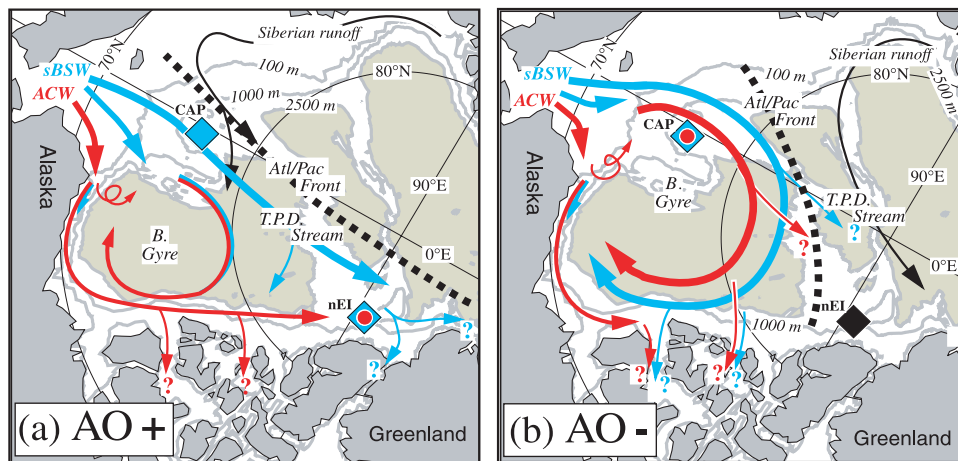


Figure 14. Schematic circulation of summer Pacific halocline water, separated into ACW (red) and sBSW (blue) components, in (a) positive Arctic Oscillation states and in (b) negative Arctic Oscillation states. Two key locations are marked by red circles (ACW) and blue diamonds (sBSW), i.e., the Chukchi Abyssal Plain (CAP) and the area north of Ellesmere Island (nEI). These are color-coded by the influence of ACW and sBSW in different regimes. The black diamond at nEI in AO- conditions indicates a lack of recent renewal of summer Pacific halocline water, at least as determined by a temperature maximum. The hypothesized outflow to the North Atlantic Ocean through the Canadian Archipelago and Fram Strait is illustrated by arrows with question marks, indicating a lack of data in our study. The Atlantic/Pacific front and the main axis of Siberian river runoff are also indicated.

summer Pacific halocline water in high and low AO states. In each state, ACW enters the Arctic Ocean principally through Barrow Canyon, while sBSW enters principally through Hanna/Central and Herald Canyons (although these stratified waters may decouple from the bathymetry and enter at other locations as well). We have identified two locations (the area north of Ellesmere Island and the Chukchi Abyssal Plain) discussed in the previous section where changes in summer Pacific halocline water signature are most evident, as the AO index varies between positive and negative states. As mentioned previously, we caution that these schematic circulations based on temperature and salinity alone may be subject to modification when analyses using chemical tracer data are applied to this subject [e.g., Moore *et al.*, 1983; Jones *et al.*, 1998]. In particular, the presence of a temperature maximum identifies areas with relatively recent renewal of summer Pacific halocline water, on timescales discussed in the previous section.

5.1. Positive AO Index

[50] Our data from the early 1990s provides a reasonably good picture of summer Pacific halocline water circulation in strongly positive AO states (Figure 14a). The Beaufort Gyre is small and composed mostly of ACW, at least in its southern limb. ACW enters the Beaufort Gyre most probably as eddies that spin off the coastal current north of Alaska. The Transpolar Drift Stream draws sBSW directly from the Chukchi Sea and swiftly (order 3–5 years) deposits it north of Ellesmere Island, and probably beyond into Nares and Fram Straits. Some evidence of this is provided by Jones *et al.* [2003], who find that the core of Pacific water fraction in Fram Strait lies between salinities of 32 and 33. The circulation during these years also brings sBSW to the Chukchi

Abyssal Plain and to the area north of Ellesmere Island. As always, the Transpolar Drift Stream is influenced by Siberian runoff on its Eurasian flank, although in positive AO states this water has been pushed by cyclonic winds farther eastward along the coastline before entering the deep basins. Some Siberian runoff might also diffuse across the Transpolar Drift Stream into the Beaufort Gyre.

[51] We speculate that outflow of summer Pacific halocline waters through the Canadian Arctic Archipelago may be geographically separated under this circulation regime, i.e., ACW should flow primarily out the western channels, while sBSW should flow primarily out the eastern channels. The boundary undercurrent probably plays a role in providing a mostly ACW signature north of Alaska, the Mackenzie River delta, and, possibly, the area north of Ellesmere Island.

[52] Our analysis thus indicates that the summer Pacific halocline water increase observed by Newton and Sotirin [1997] north of Ellesmere Island during the strongly positive AO state in the early 1990s had two components. A sBSW increase was fed by a shift in the Transpolar Drift Stream, which brought this water from the Northern Chukchi Sea directly across the Arctic Ocean to the area north of Ellesmere Island. A small ACW increase was possibly fed by a strengthening of the North American boundary undercurrent in response to a weakening of the Beaufort Gyre. However, our data coverage of this boundary pathway is sparse and more work is needed in this regard.

[53] One might also describe this circulation regime as one in which the “Atlantic-Pacific front” has been rotated cyclonically from its long-term mean position, in keeping with studies such as McLaughlin *et al.* [1996, 2002] and Morison *et al.* [1998]. Here we see that the North

American end of this front has rotated eastward, enough to allow significant outflow of relatively young (i.e., containing a detectable temperature maximum) sBSW through Nares and Fram Straits. At the other end of this front, summer Pacific halocline waters are confined to the region east of 180°E.

5.2. Negative AO Index

[54] Negative AO index (relative to the 1979–2001 mean) occurred in the 1980s, during which time we have few data points (except at the very end of the period). In the later 1990s, the AO index dropped from historically high positive values toward slightly positive or neutral conditions. Here, we take ocean and ice observations over these two time periods and attempt to synthesize them into a picture of summer Pacific halocline water circulation in negative AO index conditions. The actual circulation in the later 1990s was presumably intermediate between negative and positive states.

[55] In negative AO states (Figure 14b) the Beaufort Gyre expands, sweeping ACW eddies into its anticyclonic circulation and providing an ACW signature to waters in the Chukchi Abyssal Plain and perhaps in the Transpolar Drift Stream and downstream into Fram Strait. However, there is also strong recirculation within the Beaufort Gyre, which may bring both ACW and sBSW to the southern Beaufort Gyre, in addition to any transport to this region via the boundary current. The boundary current probably does not extend much farther along the continental slope than this point, given the strong westward surface stress from an enhanced Beaufort Gyre during these years.

[56] The expansion of the Beaufort Gyre allows a stagnation point to build north of Ellesmere Island (Figure 11), where ice and presumably upper ocean waters are deflected eastward into Fram Strait, and westward into the recirculation of the Beaufort Gyre. The ice buoy motion (Figure 12) indicates that if summer Pacific waters make it to the area north of Ellesmere Island via the Transpolar Drift Stream, they would be relatively old and thus may have lost their temperature maximum signal. This was the case in the sBSW signal (Figure 13). Other tracers may have a longer timescale for dissipation (such as the nutrient signal discussed by Moore *et al.* [1983]) and thus might persist in this region (and in its outflows to the North Atlantic Ocean) for longer than the temperature maximum.

[57] One might also describe this circulation regime as one in which the “Atlantic-Pacific front” has been rotated anticyclonically, relative to the positive AO state. Here, we see that the North American end of this front has rotated westward, suppressing the renewal of summer Pacific halocline waters north of Ellesmere Island. This is not to say that there is no summer Pacific halocline water leaving the Arctic Ocean through Nares or Fram Straits. However, what does flow out will be relatively older, and thus will have a reduced temperature maximum. Eventually, in a persistent negative AO state, the Pacific water contribution to these outflows should decrease substantially. This is supported by the chemical tracer analysis of Taylor *et al.* [2003], who found reduced Pacific water contribution in the East Greenland Current

in 1987, relative to 1998. During AO– years, outflow through the western Canadian Archipelago may be more uniform in its sBSW vs. ACW mixtures, relative to AO+ states.

6. Three Types of Pacific Halocline Water in the Arctic Ocean

[58] We have described two types of summer Pacific halocline water (ACW and sBSW) found in the Canadian Basin of the Arctic Ocean. Below these waters is typically found another water mass, with a near-freezing temperature minimum and nutrient maximum at $S = 33.1$ that is thought to originate in the Bering and/or Chukchi Sea during winter [e.g., Jones and Anderson, 1986; Cooper *et al.*, 1997]. This is often referred to as winter Bering Sea Water [e.g., Coachman and Barnes, 1961] or as Upper Halocline Water (to distinguish it from Lower Halocline Water that originates on the Eurasian arctic shelves [Jones and Anderson, 1986]).

[59] While this study is not particularly focused on winter Bering Sea Water (hereafter, wBSW), we here discuss its spatial distribution in order to propose a complete description of Pacific halocline water circulation in the Arctic Ocean. Our review of the CTD data indicates the widespread presence of a temperature minimum at $S = 33.1$ within the Canadian Basin at a depth of about 150 m (e.g., Figures 1 and 3). This is found when the overlying summer Pacific halocline water is of ACW or sBSW origin, or when it is composed of both water masses. This wide coverage in the Arctic Ocean indicates that its source may be broadly distributed over the northern Chukchi Sea shelf break. However, we cannot rule out a more localized source (e.g., in a particular Chukchi canyon) combined with enhanced mixing and dispersal in the winter.

[60] Figure 15 illustrates three schematic wintertime temperature profiles found in the Canadian Basin of the Arctic Ocean. Each profile has a relatively fresh surface mixed layer that results from ambient ocean water mixing with river discharge, plus the effects of sea ice melt and growth. Below this lies the Pacific water. Near the outer boundary of Pacific influence in the Transpolar Drift Stream (Figure 15a), sBSW overlies wBSW. In the southern Beaufort Gyre (i.e., within a few hundred kilometers of Alaska and the Mackenzie River delta, Figure 15b) ACW overlies wBSW. Between these two regions (denoted “northern Beaufort Gyre” in Figure 15c), both ACW and sBSW exist and together, overlie the wBSW. This double temperature maximum might form as a result of lateral mixing in the deep Canadian Basin (as illustrated in Figure 15c) or perhaps by direct injection off some sector of the northern Chukchi shelf where ACW and sBSW might coexist. As discussed in section 5, the geographic separation within the Canadian Basin into these three types of profiles should be most distinct during AO positive states. Finally, below the Pacific halocline waters lie waters that have their origins in the eastern Arctic Ocean and the Nordic Seas, i.e., the lower halocline and Atlantic waters. Deeper waters are not shown in this figure.

[61] The exact shape of the complex summer Pacific halocline water temperature maximum in the northern

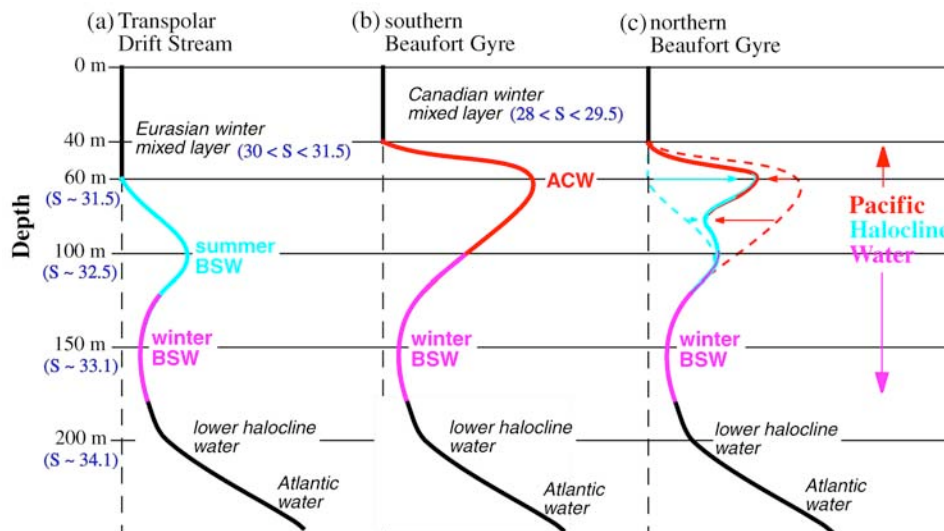


Figure 15. Schematic winter temperature profiles in the Canadian Basin of the Arctic Ocean. Pacific halocline water occupies a layer about 110–130-m thick, composed of a combination of Alaskan Coastal Water (ACW), summer Bering Sea Water (sBSW), and wBSW. The exact mix of these waters varies with position within the Basin.

Beaufort Gyre (Figure 15c) depends on the details of ACW and sBSW characteristics (temperature and depth). For example, Figure 1 shows a case where ACW overlies sBSW, although the latter does not show a particularly distinct temperature maximum. In contrast, Figure 3 shows a case where a distinct temperature minimum separates the ACW and sBSW. Could this temperature minimum itself be a distinct water mass, for example, a type of “winter ACW?” If this were the case, we might expect to find it under “summer ACW” in the southern Beaufort Gyre. It would presumably appear as a temperature minimum at a significantly lower salinity than the $S = 33.1$ typical of wBSW. However, we could find no such temperature minimum. Recent work using a combination of chemical and physical oceanographic data indicates the possibility of a saltier variety of winter ACW that might form in coastal Alaskan polynyas during years with low AO index (M. Itoh, JAMSTEC, personal communication). However, the analysis of temperature and salinity data presented in this study indicates that Pacific halocline water in the Arctic Ocean consists of three types: ACW, sBSW, and wBSW.

7. Conclusions

7.1. Three Types of Pacific Halocline Water

[62] In the Arctic Ocean, our analysis indicates the presence of two types of summer Pacific halocline water (Alaskan Coastal Water, or ACW, and summer Bering Sea Water, or sBSW) and one type of winter Pacific halocline water (winter Bering Sea Water). Relatively warmer Alaskan Coastal Water (ACW) is strongly influenced by river runoff, especially from the Yukon River, and thus occupies a fresh salinity range of $31 < S < 32$. Relatively cooler summer Bering Sea Water (sBSW) does not have this runoff influence, and thus occupies a more saline range of $32 < S < 33$. Other chemical

(e.g., nutrients) or physical tracers will probably also be useful in further distinguishing these water masses and their circulation. The total thickness of the Pacific-influenced halocline layer in the Canadian Basin is approximately 130 m (Figure 15), which is over twice the sill depth (~ 50 m) of Bering Strait. This thickness is subject to temporal variability [McLaughlin *et al.*, 2002].

7.2. Connection to the Arctic Oscillation

[63] The circulation of Pacific halocline water seems strongly influenced by the surface stress forcing associated with changes in atmospheric climate modes, i.e., the Arctic Oscillation (AO). We find that in the late 1980s and early 1990s, i.e., during strongly positive AO conditions, sBSW was transported directly across the Arctic Ocean via the Transpolar Drift Stream, while ACW was entrained into the shrunk Beaufort Gyre anticyclone and the North American boundary current (Figures 9 and 14). We speculate that this might lead to a separation of these outflows to the North Atlantic Ocean, i.e., ACW flowing through the western Canadian Arctic Archipelago, and sBSW flowing through Nares and Fram Straits. During years with lower AO index, our more limited data suggest that both types of waters are entrained into an expanded Beaufort Gyre, and the geographical separation of these water masses is reduced (Figure 14). We speculate that this should lead to a mixed ACW/sBSW signal in the Canadian Arctic Archipelago, and a weak (or old) Pacific water signature in Fram Strait.

7.3. Connection to Greenland Sea Deepwater Formation

[64] If in the future the Arctic Oscillation index is more frequently in its positive index state [e.g., Fyfe *et al.*, 1999] then our study predicts a more efficient transport of Pacific origin freshwater toward Fram Strait

and the Greenland Sea (Figure 14). This might tend to suppress deepwater formation. Such a conclusion is supported by modeling studies in which convection is enhanced when Bering Strait is artificially closed [Goosse *et al.*, 1997].

7.4. The Boundary Current

[65] Observations indicate that the North American boundary undercurrent brings only a weak ACW signal (and no sBSW) to the area north of Ellesmere Island. Although a current may exist at the Ellesmere Island continental slope, the observations of Newton and Sotirin [1997] and NPEO suggest that changes in BSW properties at that location are explainable by shifts in the Transpolar Drift Stream under AO+ and AO− (or AO neutral) conditions, rather than by changes in the hydrographic properties of the boundary current itself (Figures 13 and 14). More work is needed along the pathway of the boundary current to better define these issues.

7.5. Pacific Source Variability

[66] Long-term mooring data from Bering Strait indicate the presence of interannual variability in the northward flow from the Pacific Ocean [Roach *et al.*, 1995] <http://psc.apl.washington.edu/HLD>). Farther downstream in the central and northeast Chukchi Sea, hydrographic observations also indicate variability [Weingartner *et al.*, 1998] <http://www.ims.uaf.edu/chukchi/>). As these waters flow northward off the Chukchi shelf, they might produce spatial patterns in Pacific halocline water circulation, even if the surface stress forcing over the Arctic Ocean were constant (which it is not). In this study, we have neglected this variability. Instead, we have assumed a constant input of Pacific waters, and have assumed that the observed spatial patterns and their variability are created by wind forcing associated with the Arctic Oscillation. The result is a remarkably self-consistent picture of the circulation of the upper halocline of the Arctic Ocean. Why does this work so well? Partly, the answer may lie in the fairly generous salinity ranges assigned to ACW and to sBSW, which allows for interannual variability in this parameter. With regard to heat, we have assumed that patterns in the temperature maxima of summer water reflect cooling along its pathway through the Arctic Ocean. This is undoubtedly true. However, a careful examination of Figure 10 indicates a warming of ACW over the Northwind Ridge in the later 1990s, which may be tied to variations in its source waters, i.e., in the ACW warming observed in 1996–1997 at Bering Strait <http://www.ims.uaf.edu/chukchi/>). Our attempts to find clear correlations between Bering Strait temperatures and summer Pacific halocline water temperatures within the Arctic Ocean (not shown) yielded no definitive results. This may not be surprising, given the large interannual variability in surface ocean heat fluxes that undoubtedly occur over the shallow, partly ice-covered Chukchi Sea during summer. Further, we speculate that variability in some Pacific source waters might be significantly correlated with Pacific climate modes such as ENSO and/or the Pacific Decadal Oscillation, while the downstream circulation of these waters in the Arctic Ocean might be controlled more by winds tied to the Arctic Oscillation. More work

on the interaction between arctic shelves and basins is probably required before we can fully understand these complex issues.

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