Warming and freshening of Baffin Bay, 1916–2003

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[1] Regression analysis of historical hydrographic data is used to determine changes in temperature and salinity in Baffin Bay for the time period from 1916 to 2003. We find two distinct sets of changes in the Baffin: First, areas affected by the Atlantic inflow to Baffin Bay show substantial and statistically significant warming trends. In the more than 2000 m deep basin the warming peaks at 0.11 ± 0.06°C/decade at 700 m depth below the 640 m sill depth of Davis Strait connecting Baffin Bay to the North Atlantic Ocean. A vertical heat flux divergence of 0.25 W/m^2 is required to warm Baffin Bay below 900 m by the amount observed. The required heat appears to be advected from the shelf and slope regions of the eastern Labrador Sea via Davis Strait along the west Greenland shelf break and diffuses vertically and horizontally into the deep central basin. The North Atlantic Oscillation index accounts for about 30% of the interannual variance of temperature fluctuations below the Davis Strait sill depth. Second, areas affected by Arctic inflow to Baffin Bay show a marginally significant freshening of up to about 0.086 ± 0.039 psu/decade. This freshening trend extends along the western margin of Baffin Island to Davis Strait and into the Labrador Sea. The freshening in the northern reaches of Baffin Bay is similar in size to that at its southern reaches. Temporal variability of annual composites between these locations appears to be in phase, suggesting a swift transition of Arctic waters at 76°N to Labrador Sea waters at 62°N.

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1. Introduction

[2] Significant global warming occurred over the last 50 years in both the atmosphere [*Mann et al.*, 1995] and the ocean [*Levitus et al.*, 2000]. Ocean warming is most pronounced in the North Atlantic and a linear trend accounts for 80% of the heat content variance [*Levitus et al.*, 2000]. *Øesterhus and Gammelsrød* [1999] find that the deepest waters of the Nordic Seas warmed by about 0.1°C per decade over the last 30 years while *Robertson et al.* [2002] describe warming of deep Antarctic waters by 0.12 ± 0.07°C per decade over the last 30 years also.

[3] The northern North Atlantic has also shown significant freshening attributed to the global redistribution of freshwater [*Curry et al.*, 2003]. *Houghton and Visbeck* [2002] use salinity data from the Labrador Sea since 1948 to discuss how the salinity varies on a decadal timescale. They show correlation of salinity changes with the North Atlantic Oscillation index which emphasizes that salinity fluctuations, such as the Great Salinity Anomaly [*Belkin et al.*, 1998], are partly forced by strong winds over the entire North Atlantic [*Houghton and Visbeck*, 2002]. *Dickson et al.* [2002] discuss freshening of the Labrador Sea and Icelandic Basin waters by up to 0.01 psu per decade. *Lazier* [1995] report a freshening (and cooling) of deep Labrador Sea waters.

[4] Poor understanding of the causes for the observed warming and freshening trends prevents prediction of future changes, but we do know that the observed temporal changes potentially impact deep water formation in the Greenland and Labrador Seas [*Dickson et al.*, 2003]. Deep water formation is susceptible to small changes in the hydrological cycle which can change vertical ocean density stratification [*Dickson and Brown*, 1994; *Stommel*, 1961; *Rahmstorf*, 1995, 1998]. Small but significant changes in the hydrological cycle of the Arctic Ocean are emerging, e.g., Eurasian river discharges increase by 7% from 1936 to 1999 [*Peterson et al.*, 2002] while adjacent Siberian shelves appear to become fresher as a result of excess precipitation over evaporation, riverine discharge, and excess of ice melt over ice growth [*Steele and Ermold*, 2004]. We note, however, that the climatological sampling on shelves generally does not resolve spatially complex and temporally rapidly changing river plumes [*Münchow et al.*, 1999] and thus must be interpreted with more caution than is often done in the Arctic Ocean. Nevertheless, the hydrographic changes in Baffin Bay may indicate potential climate shifts because water exiting Baffin Bay enters the Labrador Sea, one of the two or possibly three deep convection sites in the Northern Hemisphere [*Pickart et al.*, 2002, 2003].

[5] Baffin Bay is a semienclosed basin to the north of the Labrador Sea (Figure 1). Arctic Ocean water enters Baffin Bay through three passages with sill depths less than 250 m

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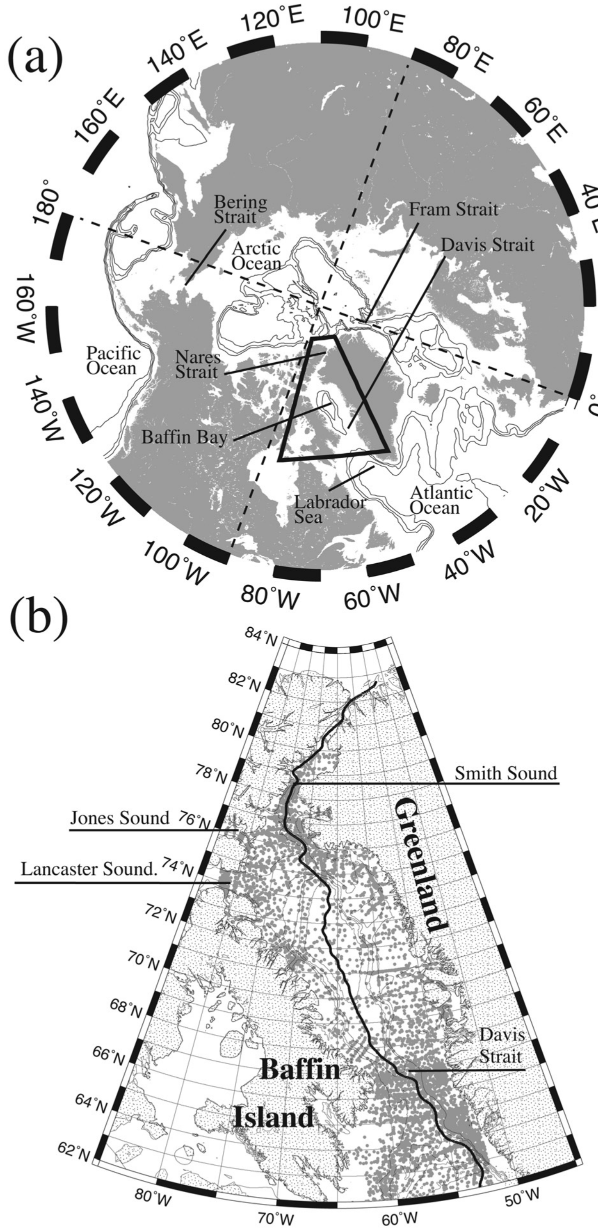

Figure 1. (a) Map of the Northern Hemisphere including the study area and (b) the data distribution in Baffin Bay, Davis Strait, and northern Labrador Sea. The thick, solid line indicates the "thalweg," that is, the deepest location at a given latitude.

at Nares Strait [*Münchow et al.*, 2006] and Smith Sound [*Melling et al.*, 2001], Jones Sound [*Melling*, 2000], and Lancaster Sound [*Sanderson and LeBlond*, 1984; *Fissel et al.*, 1982]. These cold, fresh northern inflows travel southward on the western portion of Baffin Bay as the Baffin Island Current [*Fissel et al.*, 1982], eventually exiting via Davis Strait into the Labrador Sea [*Loder et al.*, 1998; *Cuny et al.*, 2005] to become the Labrador Current [*Lazier and Wright*, 1993; *LeBlond et al.*, 1981]. Warm, saline water from the North Atlantic, mostly made of waters originating from the Irminger [*Bersch*, 1995] and East Greenland Currents [*Bourke et al.*, 1989], flows north through Davis Strait then along the eastern side of Baffin Bay as the West Greenland Current [*Smith et al.*, 1937]. The core of this flow is readily identified as a subsurface temperature maximum at about 400 m depth. It forms the eastern leg of a postulated cyclonic circulation in Baffin Bay [*Bourke et al.*, 1989; *Bâcle et al.*, 2002] that is generally used to explain observed ice distributions and water mass properties. The spatial and temporal structure, variability, or dynamics of these postulated currents are largely unobserved and are based on ice observations, ship drifts, hydrographic measurements, and the assumption of a geostrophically balanced baroclinic flow. We next briefly discuss two examples of such observations from the (largely unpublished) historical record in order to introduce spatial features whose temporal changes are the subject of this study.

[6] Figure 2 shows Baffin Bay and the location of two hydrographic sections that we use to introduce dominant water mass characteristics and their spatial distribution. Data from a 1948 section across southern Baffin Bay taken from the Danish Naval survey vessel *Hejmdal* indicates three main hydrographic features of Baffin Bay (Figure 3):

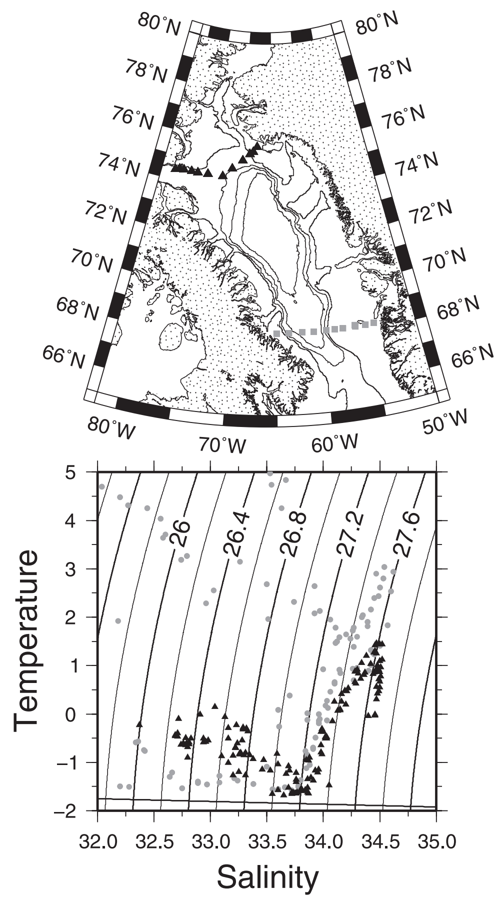

Figure 2. Map of the study area with contours of bathymetry along with the location of two sections across northern (solid symbols) and southern Baffin Bay (shaded symbols) along with their temperature salinity correlations where the contours are density anomalies (kg/m^3) and the straight line near at the bottom is the freezing line.

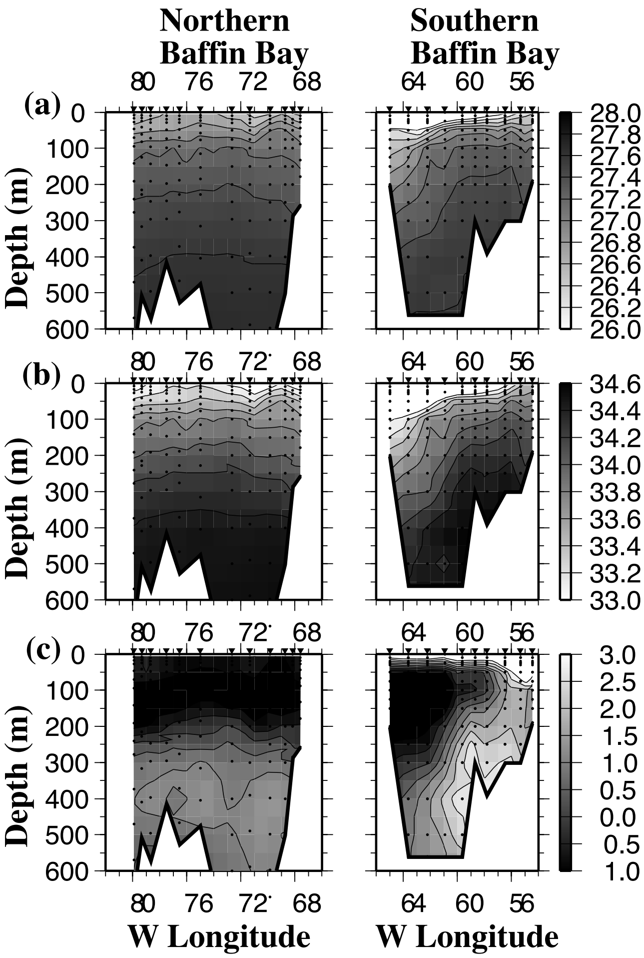

Figure 3. (a) Density anomaly (kg/m^3), (b) salinity (psu), and (c) temperature (°C) along the (left) northern and (right) southern sections as a function of longitude and depth. See Figure 2 for location and T-S properties. Inverted triangles indicate hydrographic stations while dots indicate bottle locations.

(1) warm (>+2°C) and saline (>34.4 psu) waters of the West Greenland Current below 200 m depth centered near 60°W longitude, (2) cold (>−1°C) and fresh (<34.0 psu) waters of the Baffin Island Current above 200 m depth centered near 64°W longitude, and (3) seasonally influenced surface waters that tend to be warmer and saltier on the Greenland shelf than other surface waters on this section. Isopycnals slope upward from west to east below this surface layer almost across the entire section. Such sloping isopycnals imply baroclinic pressure gradients that in a geostrophically balanced flow suggest southward surface currents relative to currents at deeper levels. This is the "classical" description of the Baffin Island Current transporting both ice, icebergs, and fresh Arctic waters southward into the Labrador Sea.

[7] The subsurface core of warmer water off Greenland, however, is not a baroclinic feature as its warm, salty waters are density compensated with colder and fresher ambient waters. The subsurface temperature maximum exceeds the freezing point by more than 3°C. It traces the entire eastern rim of Baffin Bay to 76°N and constitutes a prominent feature in temperature-salinity correlations (Figure 2, bottom). Using data from a survey of northern Baffin Bay taken from the Canadian research vessel *Labrador* in 1963, we distinguish three water masses (1) summer surface waters (warmer than 0°C, and fresher than 33.0 psu), (2) Baffin Island current waters (T < −1.0°C and S < 34.0 psu) consisting of Arctic waters, runoff, and winter surface waters, and (3) West Greenland Current waters in the lowest 2/3 of the entire section (T ~ +1.5°C and S ~ 34.5 psu) consisting of Atlantic inflow from Davis Strait. These waters are modified as they transit cyclonically through Baffin Bay. The temperature section for these data shows the warm Atlantic water, cooled to about 1°C at 400 m depth. Note the two cores over the eastern slope near 70°W and the western slope near 75°W. It appears that these waters are "wrapping around" the northern part of Baffin Bay following isobaths cyclonically.

2. Data Sources, Study Area, and Methods

[8] The U.S. National Oceanographic Data Center and the Canadian Bedford Institute of Oceanography assembled most of the hydrographic data collected over the last century in Baffin Bay and the Labrador Sea and made it available on the Web (http://www.mar.dfo-mpo.gc.ca/science/ocean/database/data_query.html). *Birch et al.* [1983] provides data location maps and references to largely unpublished reports. We also include data from a 2003 expedition to Nares Strait. The data used in this study consist of more than 740,000 data from about 9790 different profiles, spanning the time period from 1916 to 2003 between latitudes of 62°N and 80°N east of 85°W and west of 50°W longitude excluding northern Hudson Bay and Fox Basin. The data thus characterize hydrographic conditions of the northern Labrador Sea, Davis Strait, Baffin Bay, and southern Nares Strait (Figure 1b). We here use only profiles that contain estimates of both temperature and salinity. We furthermore exclude all profiles that contain salinity readings more than 35.1 psu (15 profiles). Figure 1b shows the study area and data distribution in space, while Figures 4 and 5 show the data distribution north of Davis Strait by year and month, respectively. In Baffin Bay most of the data were collected after 1950 during the summer months of July, August, and September when little ice covers Baffin Bay. Most data originate from the 1950s and 1960s when both U.S. and Canadian Coast Guards collected much data at the height of the Cold War and from the 1980s when Canadian and Danish scientists investigated the area for the potential of environmental impacts related to shipping and resource development.

[9] The quality of the salinity measurements varies. Earlier determination of salinity was done by hand using several different methods. Salinity measurements using a hygrometer, refractometer, and skilled chemist performing titrations had accuracies of 0.02 psu, 0.05 psu, and 0.01 psu, respectively [*Birch et al.*, 1983]. Since data in the early years are so scarce, their values are critical to a regression trend. Therefore we conducted sensitivity tests of the regression and error analyses, discussed below, to ensure the robustness of the trend. More specifically, we added and subtracted, in turn, 0.05 psu (the highest instrumental error) from the salinity values before 1950 in bins that already had statistically significant trends in salinity and had data from

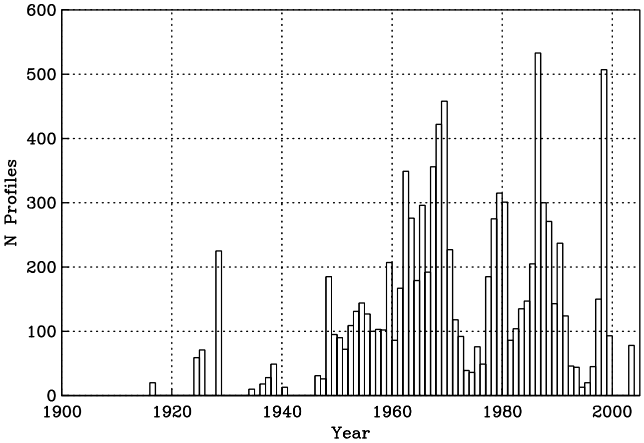

Figure 4. Distribution of data by year. Most data were gathered during the 1950s, 1960s, and 1980s.

before 1950, then recalculated the salinity trends and 95% confidence intervals. This did not change the results within the error bounds provided. Temperature records are more robust and even in 1928 absolute accuracies of 0.03 °C were well within capabilities [*Smith et al.*, 1937]. Again, sensitivity tests did not change our results within the error bounds provided [*Zweng*, 2004].

[10] The archived data contain in situ temperatures that we converted to potential temperatures. The data profiles were of two types: bottle casts and CTD casts. Most of the data from before 1970 were bottle casts, which had only a few points in each profile; most data after that time were taken using CTDs. In order to weight these data by number of profiles rather than number of data, we averaged the data within 200 m vertical bins. This minimized the CTD sampling bias in time series analyses. Likewise, some years had many more hydrographic profiles than others, hence we averaged all the profiles from each year in each bin to minimize a temporal sampling bias toward years that are well sampled. These procedures resulted in a profile-averaged and yearly averaged time series of potential temperature and salinity for each depth bin in regions of interest. We removed 163 out of 9790 profiles for which salinity and/or temperature readings below 400 m fell more than three standard deviations from the mean within four distinct regions, e.g., (1) northern Labrador Sea (62°–65°N; 67 profiles removed), (2) Davis Strait (65°–67°N; 2 profiles removed), (3) Baffin Bay (67°–76°N; 81 profiles removed), and (4) Nares Strait (76°–80° N; 13 profiles removed).

3. Baffin Bay Subsurface Warming

[11] We defined 11 nonoverlapping regions for which we construct regional averages to facilitate physically and statistically meaningful analyses. These regions are defined by isobaths and in Baffin Bay we distinguish a shelf region between the 100 m and 600 m isobaths, a slope region between the 600 m and 2000 m isobaths, and a deep basin with depths greater than 2000 m. We also separate Baffin Bay into an eastern (Greenland) and a western (Baffin Island) region that is defined by the "thalweg", that is, the location of greatest depth along a line of constant latitude (Figure 2). The Smith Sound area is bounded by Ellesmere Island at the west, Greenland at the east, 77.5°N at the south, and 79°N at the north. The Davis Strait area was bounded at the 100 m isobath to 58°W from both the east and west. In the Labrador Sea we define an eastern shelf/slope region between the 100 m and 1000 m isobath off Greenland and a deep Labrador basin with depths greater than 1000 m. These regional separations are somewhat arbitrary, but they do reflect the cyclonic circulation with inflows and outflows of waters with different properties as well as the importance of seafloor topography.

[12] For each of these regions we calculate annual averages within discrete vertical bins to conduct regression analysis, that is, we fit the "model" $T(t, z_j) = a(z_j) + b(z_j)*t$ to the data $T(t_i, z_j)$ where t is the year of the annual average, z_j is the vertical bin, T is temperature (or salinity), $a(z_j)$ is a constant offset to be determined from the data via least squares, and $b(z_j)$ is the time rate of change of temperature (or salinity) to be determined by the method of least squares also. Hence for temperature $b(z_j) > 0$ represents a linear warming trend while for salinity $b(z_j) < 0$ represents a freshening trend over the period of observations. For each regression we provide estimates of 95% confidence intervals following *Fofonoff and Bryden* [1975], assuming the degrees of freedom to be N-2 where N is the number of years with data in each time series. We thus assume a temporal decorrelation timescale of 1 year and a spatial decorrelation scale much larger than each region. Above the Davis Strait sill depth of 640 m the 1 year correlation scales appears reasonable, however, at depths below 1200 m, say, it is probably an underestimate that would bias confidence limits to be lower than appropriate on account of longer decorrelation timescales. In contrast, the presence of eddies and/or smaller than region-scale horizontal mixing and circulation features would bias our confidence limits to be higher as decorrelation scales become shorter.

[13] The warming of Baffin Bay waters is most pronounced over the deep basin where water depths reach 2400 m. Figure 6 shows the time rate of change of temperature and salinity as a function of depth along with their 95% confidence levels. It also shows the evolution of

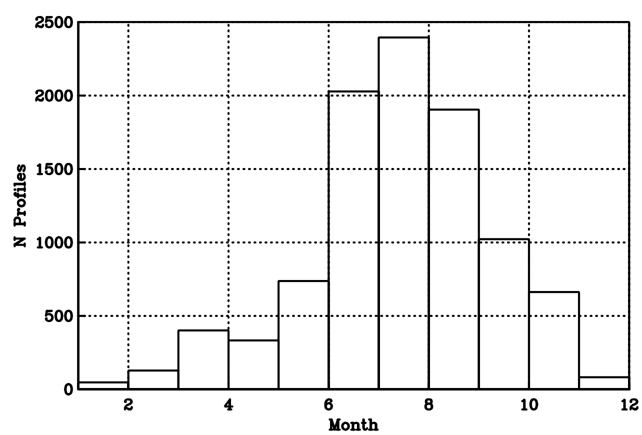

Figure 5. Distribution of data by month. The majority of data originate from the summer months July, August, and September.

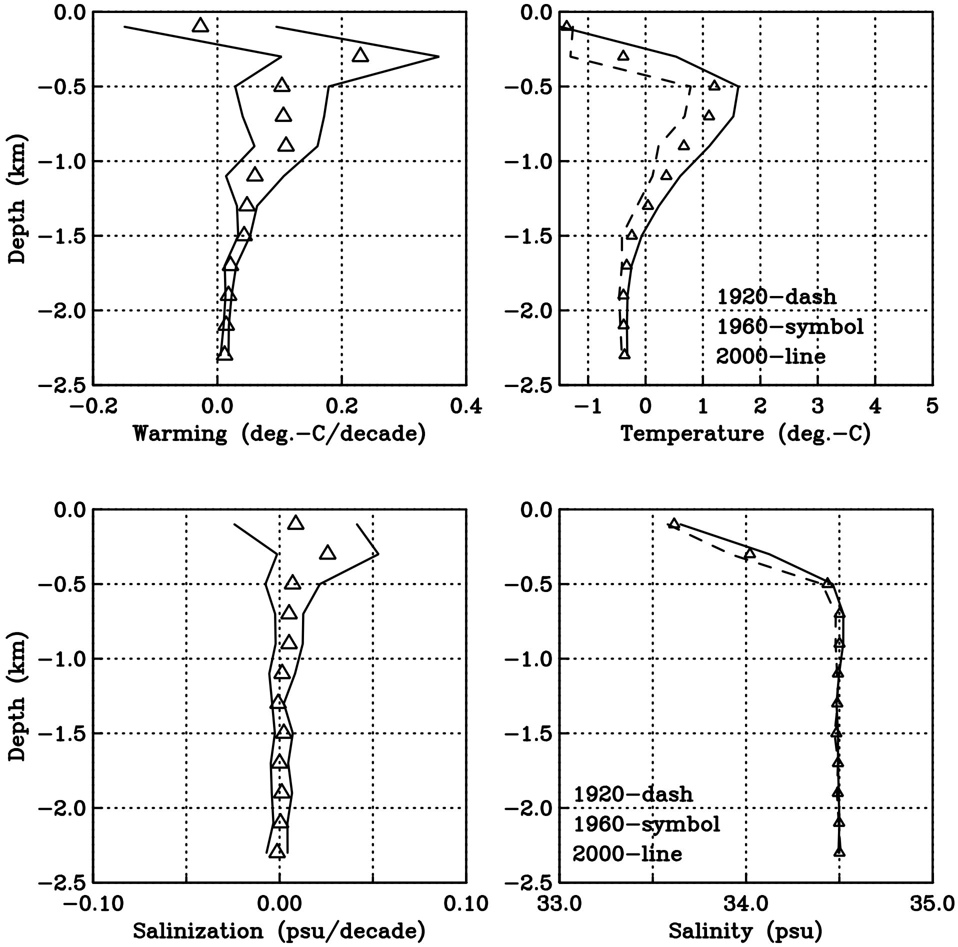

Figure 6. Time rate of change of temperature and salinity as a function of depth in Deep Basin area (H > 2000 m). Solid lines indicate the 95% percent confidence interval. Significant warming occurs throughout the water column. Also shown are predicted values in 1920, 1960, and 2000 that we obtain from the linear trend analysis.

vertical profiles for the years 1920, 1960, and 2000 predicted by the regressions. A warming over deep Baffin Bay is statistically significant from 200 m to 2400 m depth. It reaches 0.23 ± 0.13 °C/decade near 300 m depth which lays above both the sill depth of Davis Strait and above the vertical location of the subsurface temperature maximum which generally occurs near 500 m depth (Figure 2). Seasonal influences are small at these large depths thus minimizing the bias due to summer-only sampling. Below 1500 m the warming becomes smaller approaching an almost constant value of 0.03 ± 0.015 °C/decade. Within 95% confidence the warming of the entire intermediate layer from 200 to 1200 m is vertically uniform at about 0.11°C/decade while the warming of waters below 1400 m is vertically uniform at about 0.03°C/decade within the 95% confidence limits. Figure 7 provides a more direct visualization of the observed warming trend and year-to-year variations of the annually averaged temperatures. It shows the regression lines and the temperatures values from which the trends are calculated. The trends at the 50–200 m bin are statistically indistinguishable from zero while the 200–400 m bin is significant even though temperatures vary substantially from year to year. This contrasts with a very smooth and linear warming trend at the 2000–2200 m bin where temperatures exhibit little interannual variability.

4. West Greenland Subsurface Warming

[14] Figure 8 shows the time rate of change of temperature and salinity on the west Greenland shelf break within Baffin Bay. The warming trend is most pronounced in the 600 to 1000 m interval just below the temperature maximum of about +1.6°C at 500 m depth. The largest warming of about 0.15 ± 0.08°C/decade occurs in this layer. This warming is similar to the increase in temperature in the deep basin area discussed above. Note also that there is a barely significant increase in salinity of about 0.07 ± 0.05 psu/decade, that is, the slope waters near 900 m depth off west Greenland have become both warmer and saltier between 1928 and 2003. This is clear evidence of an increased contribution of West Greenland Current waters from south of Davis Strait to the changing water properties of deep Baffin Bay waters. This finding also provides evidence to support the speculation of *Bourke et al.* [1989] that the West Greenland Current enters Baffin Bay and is constrained by the 500 m isobath. Furthermore, similar changes are also observed over the western slope (600–2000 m isobaths) off

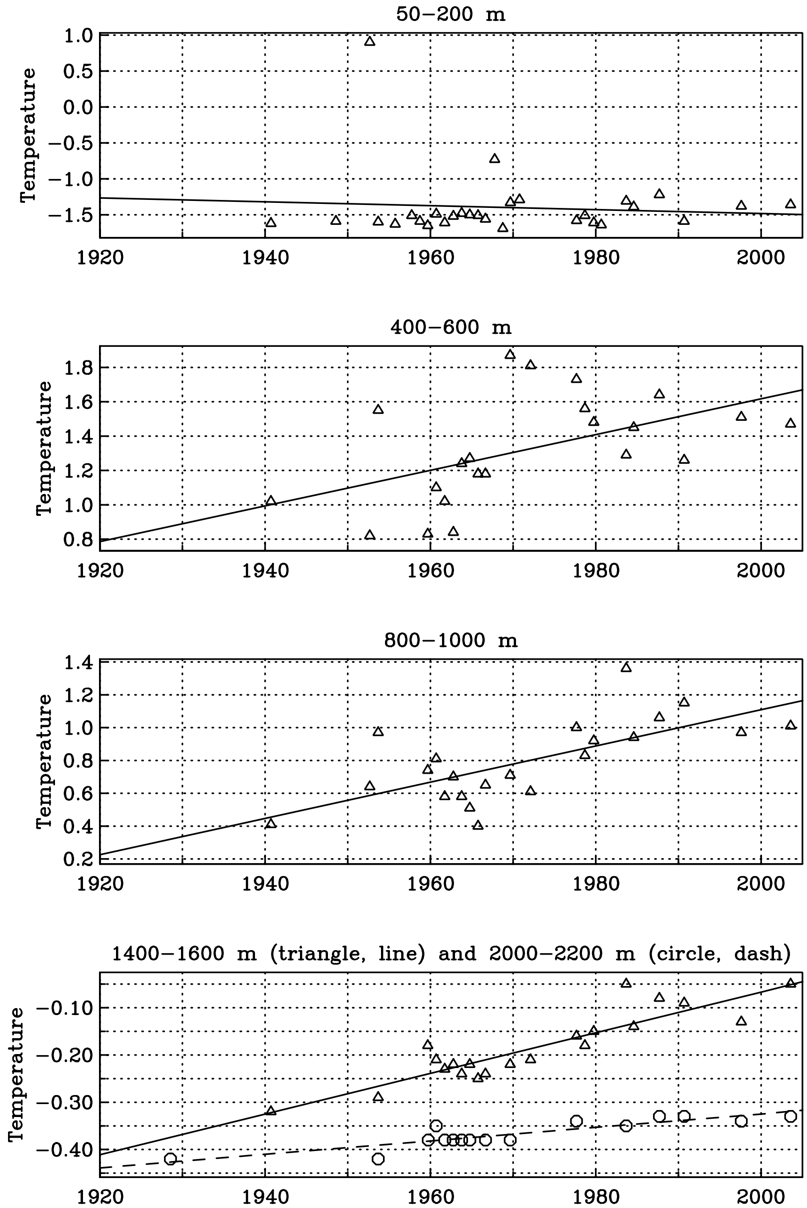

Figure 7. Time series of temperature in Deep Basin in deep Baffin Bay (defined by bottom depth H > 2000 m). Data and linear trends for five different bins are shown.

Baffin Island albeit at slightly different temperature and salinity values (Figure 9). Here we find a temperature maximum of about +1.0°C at 500 m depth and a maximum warming of 0.17 ± 0.05°C/decade at 900 m depth. Salinity increases at this depth also, by about 0.07 ± 0.06 psu/decade. Hence the influence of the West Greenland Current to the south of Davis Strait, propagates cyclonically around the rim of Baffin Bay and thus impacts the waters over the continental slope off Baffin Island at depth.

[15] It is tempting to speculate that the warming of the deeper Baffin Bay and west Greenland slope waters has its origin in the warmer and saltier waters of the West Greenland Current passing Davis Strait above the 640 m sill depth. In order to test this hypothesis we inquire about the warming trends along west Greenland to south of Davis Strait between the 600 and 2000 m isobath as well as in Davis Strait itself.

[16] The data from Davis Strait between 65°N and 67°N latitude east of 58°W longitude indicate a barely significant warming at 400–600 m depth of about 0.10 ± 0.09°C/decade. There are no statistically significant warming trends detectable over the western portion of Davis Strait (Table 1). In contrast, the narrow shelf and slope region off Greenland south of 63°N latitude inshore of the 1000 m isobath has warmed substantially, about 0.16 ± 0.10°C/decade from 1925 through 1999 (Table 2).

[17] These trends represent variability at periods longer than the record of observations, about 70 years. These observations strengthen, but do not prove, the hypothesis that the warming trends of deep Baffin Bay waters are caused by changes in southern source waters on the shelf

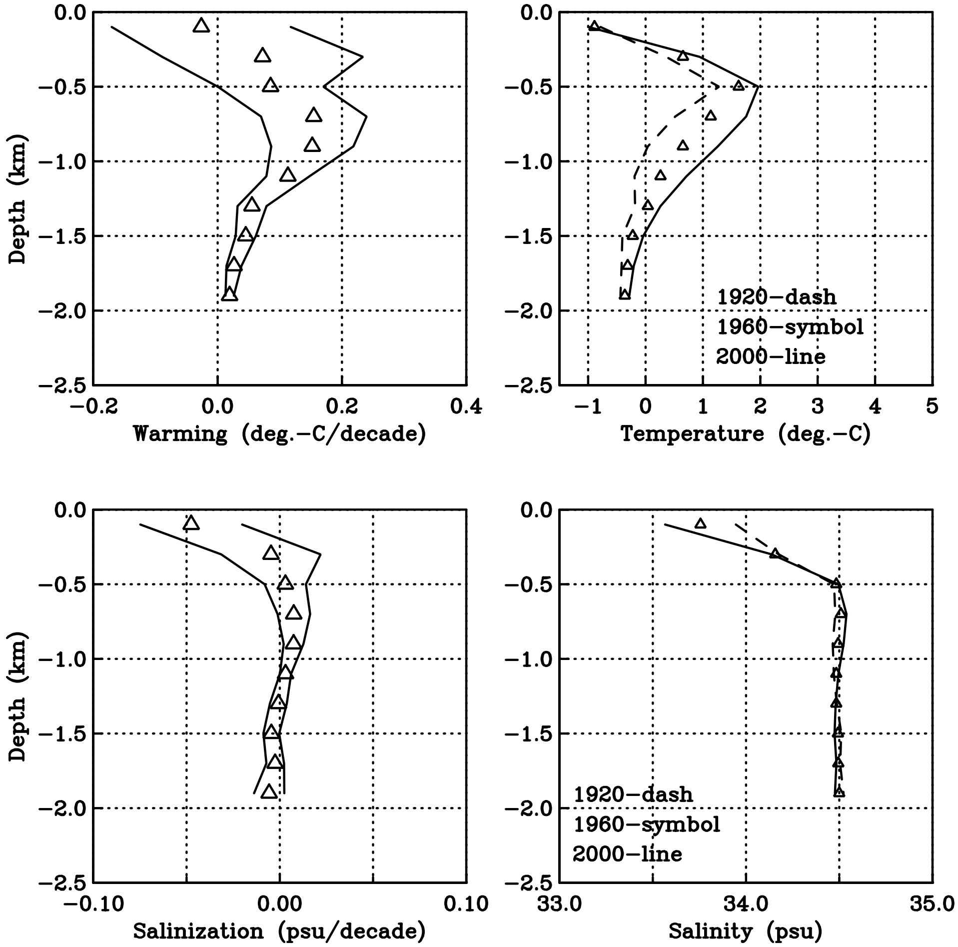

Figure 8. Time rate of change of temperature and salinity as a function of depth on west Greenland shelf break (defined by bottom depth 2000 m > H > 600 m). Also shown are predictions for the years 1920, 1960, and 2000.

and slope off southern Greenland. It is also possible that properties of the inflow from the Arctic Ocean changed in either water mass properties and/or fluxes.

5. Baffin Bay Surface Freshening

[18] At high latitudes temperature fluctuations contribute little to density fluctuations and thus vertical stratification. In contrast, spatial and temporal salinity changes are almost always synonymous with density changes and thus potentially impact the dynamics. In our analyses we find statistically significant subsurface salinity trends only north of 77°N and south of 64°N latitudes (Table 1). Waters at the northern extreme of Baffin Bay, e.g., Smith Sound at its connection to Nares Strait and the Arctic Ocean become fresher by about 0.032 ± 0.021 psu/decade in the 400–600 m depth bin during the period from 1928 to 2003 (Table 1). Waters of the deep Labrador Sea just south of Davis Strait but north of Hudson Strait become fresher also (Figure 10) by an amount ranging from 0.046 ± 0.031 psu/decade at the surface (50–200 m depth bin; Table 2) to 0.011 ± 0.004 psu/decade at depth (1200–1400 m depth bin). *Dickson et al.* [2002] report similar freshening further south in the central Labrador Sea at even greater depths. We note, however, that the waters over the eastern shelf and slope region off Greenland in the Labrador Sea north of 61°N at the 400–600 m depth bin actually become saltier by about 0.023 ± 0.013 psu/decade. Hence substantial spatial (and temporal) variability in salinity and salinity changes exist across the dynamically very active west Greenland shelf and slope regions.

[19] A systematic spatial pattern of significant freshening trends exists only over the shelf and slope regions of Baffin Bay in the surface 50–200 m depth bin which we depict in Figure 11 and Table 2. We find the largest freshening on the Baffin Island shelf, defined here from the 100 m to the 600 m isobath, of about −0.086 ± 0.039 psu/decade. The trend diminishes to −0.066 ± 0.053 and −0.048 ± 0.027 psu/decade over the continental slopes, defined from the 600 m to the 2000 m isobaths off Baffin Island. Given the different record lengths and sampling locations that all reveal a similar freshening trend from 78°N to 62°N latitude adjacent to the North American continent, we speculate that this freshening results from a common, northern source, that is, an enhanced fresh flux from the Arctic Ocean into Nares Strait, Baffin Bay, and to the Labrador Sea. We must note, however, that the surface waters of the Greenland shelf and slope regions to the north, but not to the south of Davis Strait have somewhat similar freshening trends of about

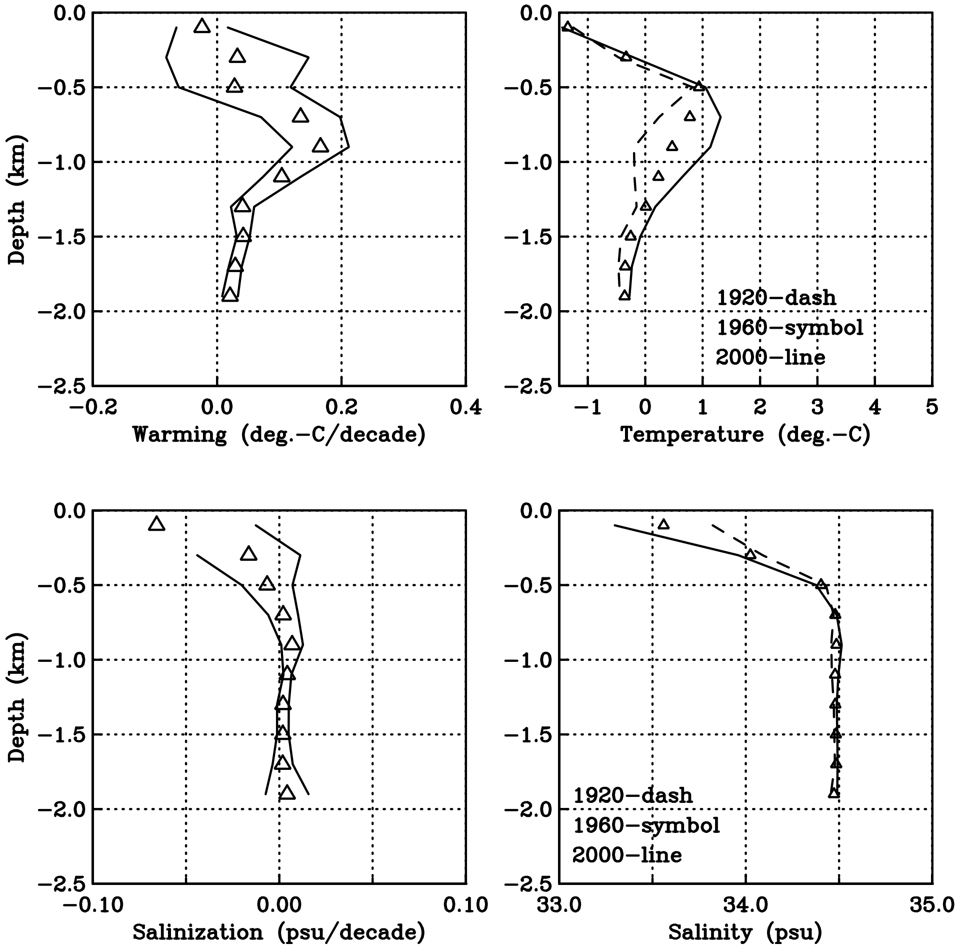

Figure 9. Time rate of change of temperature and salinity as a function of depth over the Baffin Island shelf break (defined by bottom depth 2000 m > H > 600 m). The warming is significant from 600 to 200 m with a maximum value of about 0.13°C/decade.

−0.04 psu/decade. We further speculate that increasing freshwater runoff from Greenland may contribute to the overall freshening of surface waters around the eastern rim of Baffin Bay as well. Our statements here are tentative for the potential of a sampling bias due to seasonal salinity changes exists. Recall that most observations are taken during the summer months only (Figure 5).

[20] In order to inquire about the possible salinity sampling bias, we present a 1987–1990 mooring record of salinity at 150 and 300 m nominal depth in Davis Strait [*Tang et al.*, 2006]. Figure 12 shows low-pass-filtered salinity time series with the mean values removed. The Lanczos raised cosine filter has a cutoff period near 30 days. We compare these filtered data against predictions from a least squares fit of the data to two sinusoidal oscillations to represent an annual (Sa) and a semiannual (Ssa) constituent, e.g., $S(t) = S_0 + s_1\cos(\omega_1 - \phi_1) + s_2\cos(\omega_2 - \phi_2)$ where s_1, ω_1, and ϕ_1 are amplitude, frequency, and phase of constituent 1 (Sa) while s_2, ω_2, and ϕ_2 are amplitude, frequency, and phase of constituent 2 (Ssa), S(t) are salinity predictions, and S_0 constitutes a fitted mean salinity. Figure 12a shows data from a 150 m mooring labeled E-150 that is embedded in the northward flowing West Greenland Current. Annual mean salinities are about 34.17 (1987/1988),

Table 1. Summary of Subsurface Warming Trends[a]

Location	Depth, m	Warming, °C/decade	Freshening, psu/decade	Period	N, Years	Figure
Smith Sound	400–600	0.01 ± 0.03	**−0.032 ± 0.021**	1928–2003	18	n/a
Baffin Bay, Deep	600–800	**0.11 ± 0.07**	0.005 ± 0.008	1940–2003	22	Figure 7
Greenland, Slope	600–800	**0.15 ± 0.08**	0.008 ± 0.009	1928–2003	23	Figure 8
Baffin Island Slope	600–800	**0.13 ± 0.06**	0.002 ± 0.008	1928–2003	25	Figure 9
Davis Strait, East	400–600	**0.10 ± 0.09**	0.007 ± 0.012	1924–1999	58	n/a
Labrador Sea, East	400–600	**0.16 ± 0.10**	**0.023 ± 0.013**	1925–1999	34	n/a
Labrador Sea, Deep	400–600	−0.00 ± 0.04	**−0.010 ± 0.007**	1924–1999	60	Figure 10
Davis Strait, West	400–600	0.01 ± 0.14	0.007 ± 0.022	1924–1997	31	n/a

[a]Bold values are significant at the 95% level of confidence.

Table 2. Summary of Surface Freshening Trends[a]

Location	Depth, m	Warming, °C/decade	Freshening, psu/decade	Period	N, Years	Figure
Smith Sound	50–200	**0.04 ± 0.03**	**−0.034 ± 0.033**	1928–2003	22	n/a
Baffin Bay, Deep	50–200	−0.03 ± 0.12	0.009 ± 0.033	1940–2003	29	Figure 7
Greenland Slope	50–200	−0.03 ± 0.14	**−0.048 ± 0.027**	1928–2003	32	Figure 8
Greenland, Shelf	50–200	−0.00 ± 0.07	**−0.041 ± 0.016**	1916–1999	45	n/a
Baffin Island, Slope	50–200	0.02 ± 0.04	**−0.066 ± 0.053**	1940–2003	33	Figure 9
Baffin Island, Shelf	50–200	−0.00 ± 0.07	**−0.086 ± 0.039**	1924–2003	27	n/a
Davis Strait, East	50–200	**−0.12 ± 0.11**	**−0.026 ± 0.020**	1924–1999	60	n/a
Labrador Sea, East	50–200	0.10 ± 0.12	−0.009 ± 0.032	1924–1999	60	n/a
Labrador Sea, Deep	50–200	−0.08 ± 0.13	**−0.046 ± 0.031**	1924–1999	60	Figure 10
Davis Strait, West	50–200	0.04 ± 0.14	**−0.046 ± 0.036**	1924–1997	34	n/a

[a]Bold values are significant at the 95% level of confidence.

34.35 (1988/1989), and 34.33 (1989/1990) psu. The "seasonal" signal is a composite of the semiannual ($s_2 = 0.09$ psu amplitude) and annual ($s_1 = 0.13$ psu amplitude) constituent. Nevertheless, much larger variability exist at nonsolar periods ranging from monthly (winter 1988/1989) to 3 month (1987/1988) variations. This nondeterministic behavior contrasts dramatically with signals depicted for a 150 m and a 300 m mooring closer to Baffin Island labeled B-150 and B-300, respectively. These moorings are embedded in the southward flowing Baffin Island Current. Annual mean salinities are 33.46 psu for the 150 m record B-150 (1987/1988) and 34.08 (1987/1988), 34.25 (1988/1989), and 33.85 (1989/1990) for the 300 m record B-300. These fresher waters exhibit a much stronger and persistent seasonal cycle that is dominated by the annual constituent with amplitudes of $s_1 = 0.22$ and 0.17 psu for instruments moored at 150 m and 300 m depths, respectively. The nondeterministic component of the signal is much smaller than that for the mooring off Greenland. Hence we conclude, that some of the salinity trends reported above for the Baffin Island surface waters could have larger errors than presented above on account of a biased sampling in the

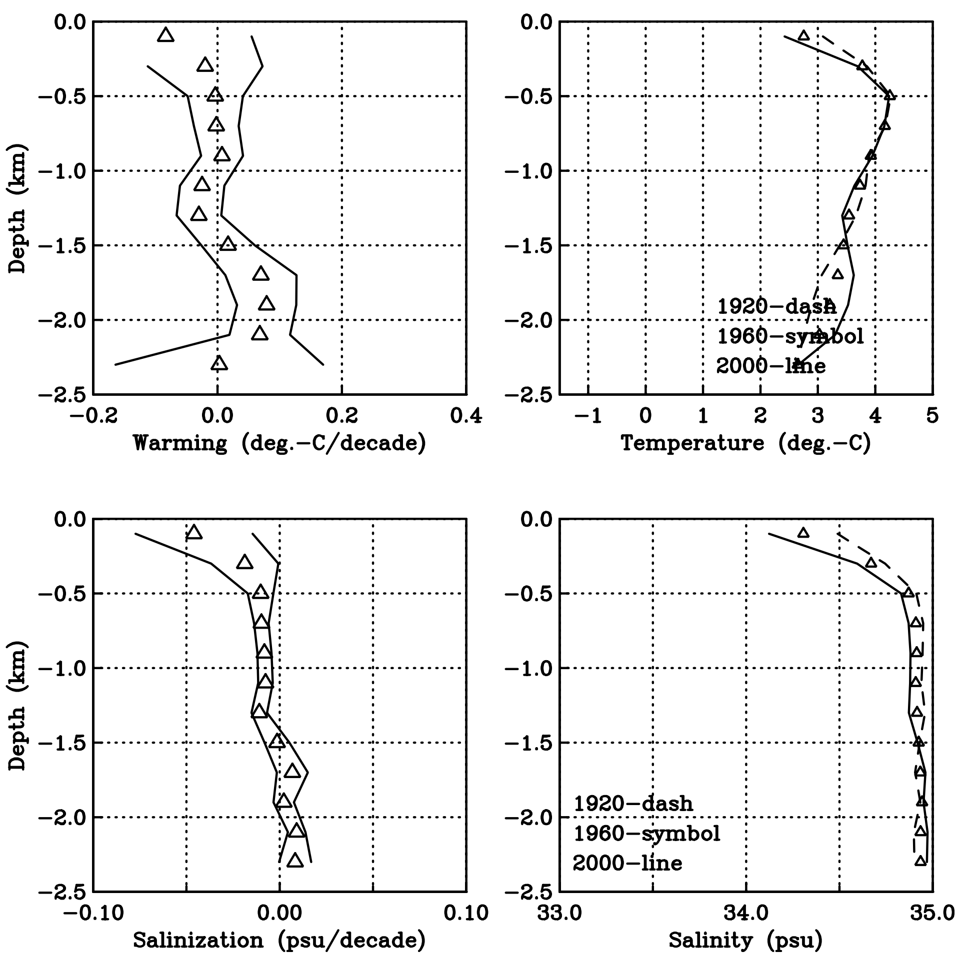

Figure 10. Time rate of change of temperature and salinity over the deep (H > 1000 m) Labrador Sea north of 61°N. Note the significant freshening (negative salinization) from the surface to about 1500 m depth.

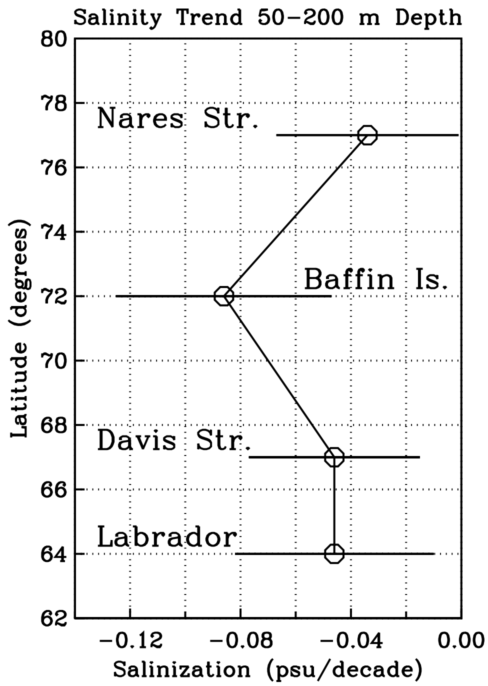

Figure 11. Salinity trends along Baffin Island from Nares Strait in the north to the northern Labrador Sea in the south. Error bars are 95% confidence limits (Table 2).

summer (July–October) when seasonal salinity variations are large off Baffin Island (e.g., Figure 12c). Note also that the peculiar phasing of Sa and Ssa results in large seasonal salinity variations during the summer months, that is, our trend analyses will not improve if we limit it to data from summer months. Unfortunately, we cannot make a more definite statement than that the 1987–1990 mooring data from Davis Strait suggests a seasonal cycle of salinity that, we speculate, varies on the scale of the internal Rossby radius of deformation. The latter is not resolved by the mooring data, furthermore, mooring data are limited to depths below 150 m depth even though most freshwater flux resides within 100 m of the surface [*Münchow et al.*, 2006].

6. Discussion

[21] We propose that the deep warming is facilitated by heating from above by a layer near the Davis Strait sill depth of about 600 m. Testing the feasibility of this argument, we calculate the amount of heat needed to warm the deep waters of Baffin Bay. The heat content H in J/m^2 of a layer between two depth levels $D1$ and $D2$ is

$$H = \int_{D1}^{D2} \rho C_p (\theta - \theta_0) dz$$

[*Send et al.*, 1987], here ρ is the density (1026 kg/m^3) and C_p is the specific heat capacity of sea water (3986 J/kg/°C), θ and θ_0 are potential and reference temperature, respectively. Assuming that temperature is constant horizontally within each vertical layer, integrating over the horizontal area A, and taking a derivative with respect to time, we get (in a discrete form):

$$A\Delta H/\Delta t = \rho C_p V \Delta\theta/\Delta t$$

where $\Delta H/\Delta t$ is the rate of change of heat content with time (W/m^2), $\Delta\theta/\Delta t$ is the rate of change of potential temperature with time (~0.1°C per decade), and $V = (D2 - D1)*A$ Is the volume (41,700 km^3) of a 200 m thick layer extending over the area of Baffin Bay below 900 m (208,500 km^2). In this area, the assumption of spatially constant temperature is fairly good as the temperatures and salinities are largely uniform at this depth. With these scales we find that it requires about 27 × 10^9 Watts to warm deep Baffin Bay by the observed amount. In the absence of advection this value corresponds, if spread uniformly over area A, to a vertical heat flux of about 0.25 W/m^2. While this number is small relative to sea surface heat fluxes of about 100 W/m^2, it is large relative to geothermal heating of about 0.05 W/m^2 [*Adcroft et al.*, 2001].

[22] From subsurface volume integrals of the heat budget equations, we find

$$\partial\theta/\partial t \cdot (V/A) = (u_1 \cdot \theta_1 + u_2 \cdot \theta_2 + u_3 \cdot \theta_3 - u_4 \cdot \theta_4) + F_G/(\rho C_p) - \kappa_z \partial \cdot \theta/\partial z$$

where local potential temperature changes $\partial\theta/\partial t$ are caused by the horizontal heat flux divergence, geothermal heating

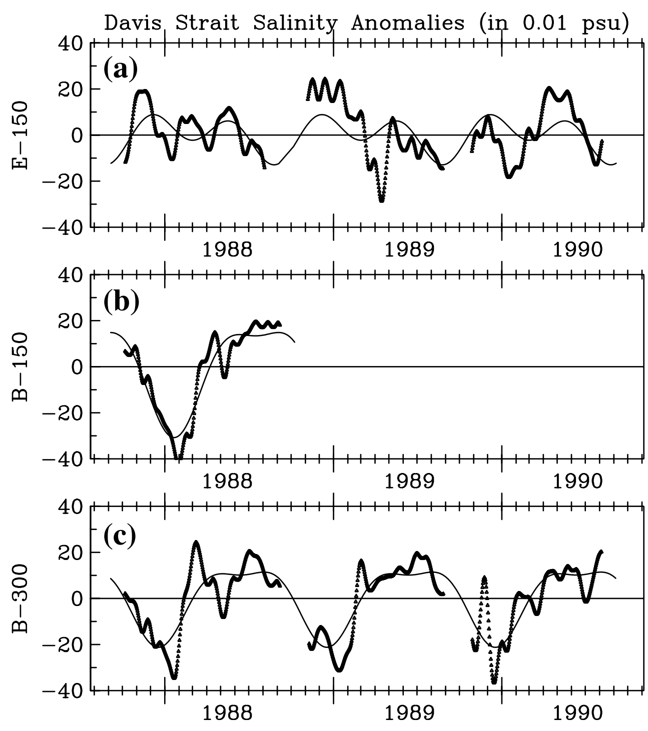

Figure 12. Time series of low-pass-filtered salinity (symbols) and predictions (lines) for annual and semiannual constituents for (a) a location over the Greenland slope at 150 m nominal depth (E-150), (b) for a location near Baffin Island at 150 m nominal depth (B-150), and (c) for the same location near Baffin Island but at 300 m nominal depth (B-300).

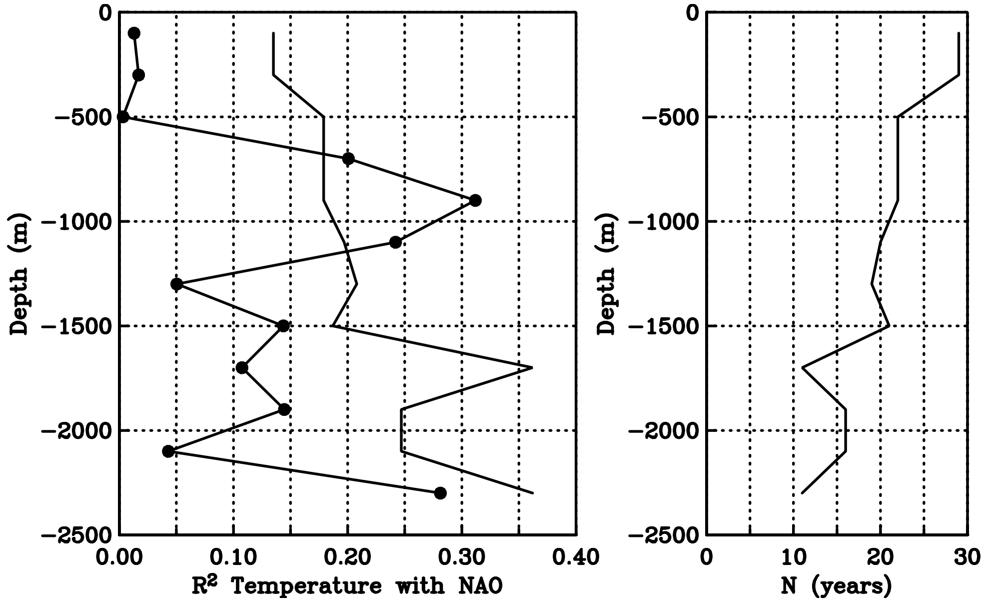

Figure 13. Correlation squared (r^2) between deep Baffin Bay temperatures lagged by 1 year (Figures 6 and 7) and the NAO as a function of depth. Significant levels at 95% are shown as solid lines without symbols. The correlations are significant at the 95% level between 600 and 1200 m.

F_G, and vertical diffusion across the upper level of the (subsurface) integration volume with a turbulent diffusivity κ_z. Ice melt, ice growth, precipitation, and evaporation all take place at the sea surface and these processes are incorporated in the water properties within a layer above our subsurface integration volume. Here A is the area of a horizontal slice at a given depth below the surface and V is the volume enclosed by this slice and the bottom [*Timmermans et al.*, 2005]. The Baffin Bay heat flux divergence consists of (1) the inflow of warm and salty West Greenland Current water from eastern Davis Strait $u_1 \cdot \theta_1$, (2) the inflow of water cold and fresh Arctic waters $u_2 \cdot \theta_2$, (3) the inflow of cold and salty brines from coastal polynyas $u_3 \cdot \theta_3$, and (4) the outflow of mixed waters through western Davis Strait $u_4 \cdot \theta_4$ where $u_1 + u_2 + u_3 - u_4 = 0$ in the absence of any other sources or sinks of volume flux. A similar expression for salt is

$$\partial S/\partial t \cdot (V/A) = (u_1 \cdot \Sigma_1 + u_2 \cdot \Sigma_2 + u_3 \cdot \Sigma_3 - u_4 \cdot \Sigma_4 - \kappa_z \cdot \partial S/\partial z.$$

These much simplified versions of the heat and salt budgets demonstrate why it is difficult to assign a definite cause to the observed subsurface heating in the absence of reliable velocity or even volume flux estimates. Nevertheless, they do allow us to estimate vertical diffusivities if we choose our upper horizontal control surface A at a depth below the Davis Strait sill depth as then the divergence terms vanish. Neglecting geothermal heating, we estimate $\kappa_z \sim 0.5 \times 10^{-5}$ m^2/s using $\partial\theta/\partial t \sim 0.01$°C/y, $\partial\theta/\partial z \sim 0.5 \times 10^{-4}$°C/m with V $\sim$ 214,000 km^3 and A $\sim$ 209,000 km^2 at 900 m depth. Repeating this calculation for salinity, we find $\kappa_z \sim 3 \times 10^{-5}$ m^2/s with $\partial S/\partial t \sim 1.6 \times 10^{-11}$ psu/s, $\partial S/\partial z \sim 0.5 \times 10^{-3}$ psu/m for 900 m depth also. These values tend toward the low end of canonical values of "pelagic diffusivity" of diapycnal ocean mixing away from boundaries [*Munk and Wunsch*, 1998]. Hence spatially uniform turbulent diffusion is sufficient to diffuse heat vertically downward across isopycnals at 900 m depth in Baffin Bay. The remaining question then becomes what causes the heat in the upper layers of Baffin Bay, e.g., above the Davis Strait sill depth, to change.

[23] In order to speculate on the causes of the observed changes in heating $\partial\theta/\partial t$ and freshening $-\partial S/\partial t$, we entertain five possibilities, that is, (1) an increased inflow u_1 from the eastern Labrador Sea, (2) a diminished inflow u_2 from the Arctic, (3) diminishing deep water formation within Baffin Bay in northern coastal polynyas, e.g., $u_3 \cdot \theta_3$ (R. D. Muench, personal communication, 2005), (4) geothermal heating, and (5) changing temperatures θ_1 of southern source waters. Other possibilities exist. We cannot address possibility 1 since the West Greenland Current hydrographic signal is density compensating (warm and salty, see Figure 3). Thus we cannot use geostrophic currents calculated from hydrographic observations because a barotropic flow component (often added as a "level of no motion") likely dominates the flow. Furthermore, the absence of winter data from northern coastal areas prevents an investigation of possibility 3. Hence we cannot conclusively favor one process over another. Nevertheless, we can and will discuss interannual variability using the North Atlantic Oscillation (NAO) index as a proxy that may indicate a connection between Baffin Bay waters and northern North Atlantic atmospheric forcing. Our finding of significant correlations will strengthen, but not prove the hypothesis that Baffin Bay warming is driven by a source on the shelf and slope off Greenland south of Davis Strait.

[24] The NAO represents the dominant mode of atmospheric variability over the northern North Atlantic Ocean without a preferred temporal scale [*Hurrell et al.*, 2003]. It relates closely to the midlatitude Azores High and the

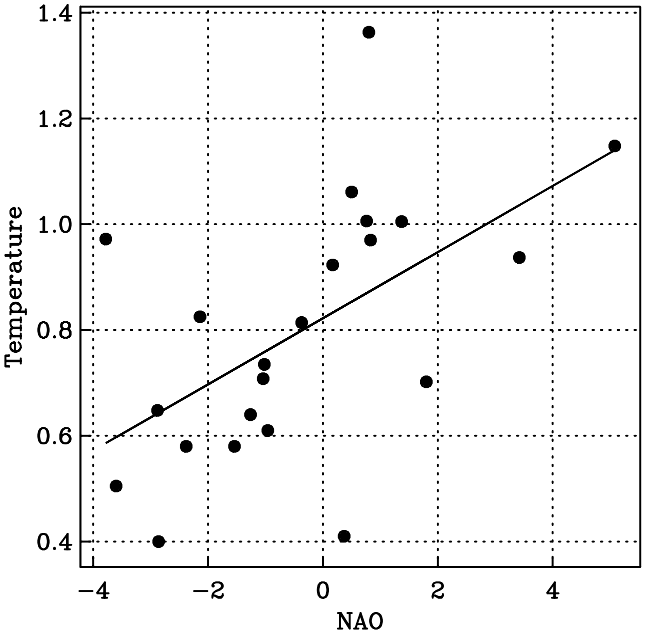

Figure 14. Scatter plot of NAO values and deep Baffin Bay temperatures at 900 m depth along with regression line. The regression indicates that temperatures at 900 m depth increase with increasing value of NAO.

subpolar Icelandic Low pressure systems with their attendant storm tracks. It dominates low-frequency atmospheric (and oceanographic) variability over much of the North Atlantic. We here use the NAO as an indicator of "upstream" surface forcing for Baffin Bay. During negative NAO phases winter storms cross the North Atlantic at a more southerly location resulting in milder, wetter winters off southern Greenland while during the positive NAO phases more storms pass at more northerly locations resulting in cooler and drier winters off Greenland. The atmospheric circulation is stronger and more cyclonic during positive NAO years in the Labrador Sea. It is not clear, however, how this relates to the oceanic circulation in Baffin Bay and near Davis Strait.

[25] We use the *Hurrell et al.* [2003] annual station-based index that represents a normalized difference of sea level pressure between the Azores and Iceland. Figure 13 shows the correlation squared (r^2) of the NAO with our deep Baffin Bay hydrographic observations as a function of depth for a 1 year lag. The temporal variations of temperatures above the 600 m sill depth of Davis Strait do not correlate with the NAO, however, between 600 and 1200 m below the surface we find correlations that are significantly different from zero at the 95% confidence level. Since a 95% confidence implies a (random) 1:20 chance of false detection, perhaps one of our 12 independent correlation estimates might be false, however, there are three significant correlations clustered about 900 m where the NAO explains between 20 and 30% of the variance of annually averaged temperatures. This is both a remarkable and robust result. Note that this is the layer below the subsurface temperature maximum (Figure 6) with large warming trends (Figure 7). The correlations are such that temperatures increase with increasing NAO (Figure 14). For example, 900 m temperatures during a strongly positive NAO year such as 1990 are about 0.5°C warmer than during a strongly negative NAO year such as 1964. This finding contradicts the interpretation of differences between ocean conditions of years with extreme NAO states as dramatic trends in climate change. Futhermore, our data contains scatter and other factors impact water temperatures besides the NAO. Nevertheless, significant warming trends in deep Baffin Bay, west Greenland, and significant correlations between subsurface temperatures and NAO all emphasize that Baffin Bay waters below the Davis Strait sill depth relate to upstream southern source waters off west Greenland.

7. Conclusions

[26] Areas of Baffin Bay associated with the warm, saline West Greenland Current are warming. The deep basin temperatures in the region with depths greater than 2000 m, exhibit a statistically significant warming from 400 to 2400 m depth. The maximum warming occurs between 600 and 800 m depths, and has a magnitude of almost 0.2°C/decade. The warming is strongest between 600 and 1200 m; below 1200 m the warming is weaker by a factor of 5, but it is still significant. On the west Greenland slope, the area between 600 and 2000 m isobath that divides the west Greenland shelf from the deep basin, warming by a similar amount is evident also. The statistically significant warming appears between 600 and 1200 m depths, with a maximum warming of almost 0.2°C/decade also in the 600–800 m depth interval. The warming in these areas is most likely caused by an increase in temperature of the inflowing Atlantic waters of the West Greenland Current. This speculation is supported by a significant correlation of subsurface temperature fluctuations with the NAO. Geothermal heating within Baffin Bay is insufficient to provide the energy needed to warm the water column by the amount observed over the last 80 years.

[27] We also find statistically significant freshening of the surface (50–200 m depths) waters especially along Baffin Island from 76°N all the way south to 62°N, however, seasonally biased summer sampling along with a strong and asymmetric seasonal signal of surface salinity quantified for Davis Strait prevents a firm conclusion. A carefully designed study of upper ocean salinity variability over multiple years is needed to fully resolve seasonal salinity cycles that could be used to "correct" for the summer sampling bias in the historical record. Such studies are presently underway as part of the Arctic-Subarctic Ocean Flux experiment.

[28] **Acknowledgments.** This study would not have been possible without the gracious Web posting of the historical data by the Bedford Institute of Oceanography. The 1997 and 2003 data were collected from aboard the CCGC *Louis St. Laurent* mastered by Captain Gomes and the USCGC *Healy* mastered by Captain Oliver, respectively, with Kelly Falkner serving as Chief Scientist during both these expeditions to Baffin Bay. She, as well as Mike Steele, Richard Garvine, Robin Muench, and an anonymous reviewer, provided detailed and constructive criticism that substantially improved this manuscript. We gratefully acknowledge financial support from the National Science Foundation (OPP-0230236).

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