2010

On the Nature of Winter Cooling and the Recent Temperature Shift on the Northern Gulf of Alaska Shelf

Markus A. Janout

Thomas J. Weingartner

Thomas C. Royer
Old Dominion University

Seth L. Danielson

Repository Citation

Janout, Markus A.; Weingartner, Thomas J.; Royer, Thomas C.; and Danielson, Seth L., "On the Nature of Winter Cooling and the Recent Temperature Shift on the Northern Gulf of Alaska Shelf" (2010). *CCPO Publications*. 240.
https://digitalcommons.odu.edu/ccpo_pubs/240

Original Publication Citation

Janout, M. A., Weingartner, T. J., Royer, T. C., & Danielson, S. L. (2010). On the nature of winter cooling and the recent temperature shift on the northern Gulf of Alaska shelf. *Journal of Geophysical Research: Oceans, 115*, C05023,. doi:10.1029/2009jc005774

JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 115, C05023, doi:10.1029/2009JC005774, 2010

Click Here for Full Article

On the nature of winter cooling and the recent temperature shift on the northern Gulf of Alaska shelf

Markus A. Janout,[1] Thomas J. Weingartner,[1] Thomas C. Royer,[2] and Seth L. Danielson[1]

Received 31 August 2009; revised 23 November 2009; accepted 28 December 2009; published 28 May 2010.

[1] In spring 2006 and 2007, northern Gulf of Alaska (GOA) shelf waters were ~1.5°C below average throughout the ~250 m deep shelf and the salinity-dependent winter stratification was anomalously weak due to above (below) average surface (bottom) salinities. Spring 2007 and 2008 temperatures were also ~−1.5°C below average, but the anomalies were confined to the upper 100 m due to moderate salt stratification. Shelf temperatures in these 2 years were among the lowest observed since the early 1970s, thus interrupting an approximately 30-year warming trend. We examined winter cooling processes using historical conductivity-temperature-depth (CTD) profiles and mooring data from hydrographic station GAK1. The 2006 and 2007 cooling was associated with anomalously strong atmospheric heat loss in November 2006 and March 2007 and below-average fall runoff, which weakened winter stratification and allowed the late cooling to penetrate throughout the water column. In 2007 and 2008, early winter cooling was weak, fall runoff large, and stratification moderate at 100 m so that spring temperature anomalies were trapped to the upper 100 m. Analysis of the 40 year GAK1 CTD record indicates that winter averaged air-sea heat flux and salinity stratification anomalies explain 81% of the variation in deep (100–250 m) GOA temperatures. Although the timing and magnitude of winter runoff influences the shelf temperature distribution, temperature anomalies are a consequence of three-dimensional circulation and mixing processes. These involve the complex, but poorly understood, interplay among the air-sea heat flux; the ocean heat flux convergences; the stabilizing influence of runoff; and the destabilizing effects of cooling, vertical mixing, and the wind-driven cross-shelf buoyancy flux.

Citation: Janout, M. A., T. J. Weingartner, T. C. Royer, and S. L. Danielson (2010), On the nature of winter cooling and the recent temperature shift on the northern Gulf of Alaska shelf, *J. Geophys. Res.*, *115*, C05023, doi:10.1029/2009JC005774.

1. Introduction

[2] The northern Gulf of Alaska (GOA) shelf is wide (~150 km) and deep (150–260 m) and supports a productive ecosystem and important commercial fisheries. Atmospheric conditions are governed by the strength and position of the Aleutian Low (AL), which results in predominantly cyclonic (downwelling-favorable) winds around the GOA [*Wilson and Overland*, 1986] and heavy coastal precipitation. The coastal circulation is governed by the wind- and freshwater-driven Alaska Coastal Current (ACC) [*Schumacher and Reed*, 1980; *Royer*, 1981; *Johnson et al.*, 1988; *Schumacher et al.*, 1989; *Stabeno et al.*, 1995, 2004; *Weingartner et al.*, 2005]. The ACC, which has remarkable seasonal and interannual variability, transports heat and freshwater along the coast from the British Columbian shelf to the northern GOA and into the Bering Sea [*Royer*, 1981; *Weingartner et al.*, 2005]. Density variations in this subpolar ocean are primarily salinity-dependent [*Royer*, 1982], especially from fall through spring, so that stratification and the onset of the spring bloom are strongly influenced by the large coastal runoff. Coastal discharge increases in spring from snowmelt and is a maximum in fall, when precipitation and runoff are greatest [*Royer*, 1982]. Nutrients are supplied to the deeper GOA shelf during summer when downwelling relaxes and are vertically mixed in winter, whereupon they are available for the spring bloom [*Childers et al.*, 2005].

[3] In spring 2007, oceanographic monitoring over the entire water column of the northern GOA shelf revealed the lowest temperatures observed in ~35 years and anomalously weak stratification. The cool signal persisted through fall 2007 and developed anew in middle to late winter 2008, although at that time the largest negative temperature anomalies were confined to the upper 50 m. This recent cooling interrupts long-term trends of increasing stratification, warming (~0.8°C in 30 years), and freshening (~0.1 salinity decrease over 30 years) [*Royer*, 2005; *Royer and Grosch*, 2006] of the

[1]Institute of Marine Science, University of Alaska, Fairbanks, Fairbanks, Alaska, USA.
[2]Center for Coastal Physical Oceanography, Old Dominion University, Norfolk, Virginia, USA.

0148-0227/10/2009JC005774

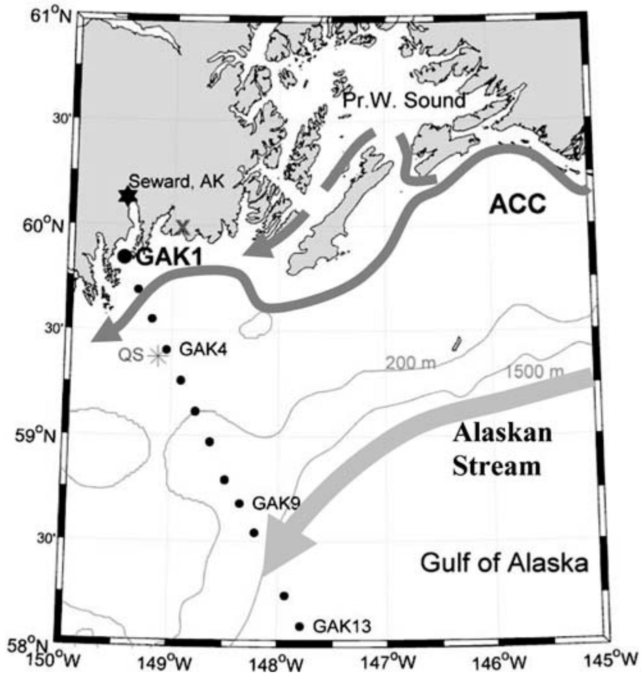

Figure 1. Map of the northern Gulf of Alaska, including GAK1 (large dot) and the Seward Line (dots). Bathymetric data from the General Bathymetric Chart of the Oceans database shows the 200 and 1500 m isobaths. Shaded lines indicate the course of the Alaska Coastal Current and the Alaskan Stream. Locations of QuikSCAT and NCEP grid points are marked by a star and a cross.

GOA's upper layer (0–100 m). Moreover, recent temperatures resemble those of the early 1970s, prior to the 1976–1977 regime shift [*Hare and Mantua*, 2000]. Herein we describe the evolution of shelf temperatures in the winters of 2006–2007 and 2007–2008 and undertake a broader review of the processes that control winter cooling on the northern GOA shelf.

[4] The paper is organized as follows. After presenting the data and the methods in section 2, we describe the recent cooling on the northern GOA shelf (section 3.1), examine the onset of the cooling and the evolution of thermohaline properties in 2006–2007 compared to other recent years (sections 3.2 and 3.3), and then place these findings in a broader climatological context (sections 3.4–3.6). Section 4 summarizes the paper.

2. Data and Methods

[5] We use hydrographic (temperature and salinity) data from along the Seward Line but primarily from station GAK1 (~59.8°N, 149.5°W), which is the innermost station on this line (Figure 1). This 263 m deep station (http://www.ims.uaf.edu/gak1/) has been occupied on a quasi-monthly basis since 1970. The record consists of ~460 profiles, although there are fewer observations from the early 1970s and 1980s (Figure 2). The Seward Line was occupied repeatedly from 1997 to 2004 as part of the NOAA/NSF-funded Northeast Pacific (NEP) Global Ocean Ecosystems Dynamics (GLOBEC) program [*Weingartner et al.*, 2002], and since 2005 with support from the North Pacific Research Board for biannual (May and September) cruises. Prior to the mid-1980s, the accuracies in temperature and salinity are ±0.02°C and ±0.05, respectively, and since then the accuracies are better than or equal to ±0.01°C and ±0.01. Prior to 1975, the salinity data originated from bottle data collected at discrete depths. We use the GAK1 conductivity-temperature-depth (CTD) data to update *Royer and Grosch*'s [2006] temperature and salinity anomaly time series after removing the warming and freshening trend per their analysis. Since 1997, the GAK1 CTD time series has been complemented by moored temperature and salinity data. The mooring was deployed annually (except in 1999 and 2003) and contains Seabird SeaCATs or MicroCATs at nominal depths of 30, 60, 100, 150, 200, and 250 m. Pre- and postcalibrations indicate that sensor drifts did not exceed ~0.01°C yr^{-1} for temperature and 0.03 yr^{-1} for salinity during any yearlong deployment.

[6] For the 1970–2008 period we use National Centers for Environmental Prediction (NCEP) reanalyzed estimates of zonal and meridional winds and air temperature at ~60°N, 149°W, the grid point nearest GAK1. In addition, we use QuikSCAT wind data at 59.375°N, 149.125°W, which is ~55 km seaward of GAK1. This location reduces the potential bias due to localized gap winds [*Macklin et al.*, 1988] on estimates of the along-shelf winds. The QuikSCAT wind estimates are from twice-daily swaths with daily averaged data available at 25 km resolution. Over the northern shelf, the along-shelf component of the wind is nearly zonal [*Royer*, 2005; *Schroeder*, 2007], with downwelling-favorable winds

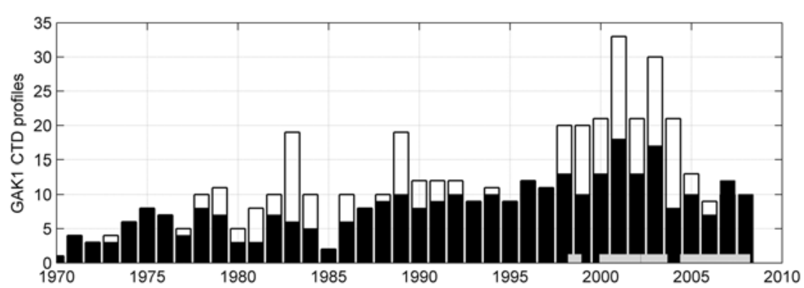

Figure 2. Annual number of recorded GAK1 CTD profiles (white) and the number of profiles used in this study (black). Gray horizontal bars indicate time of mooring coverage.

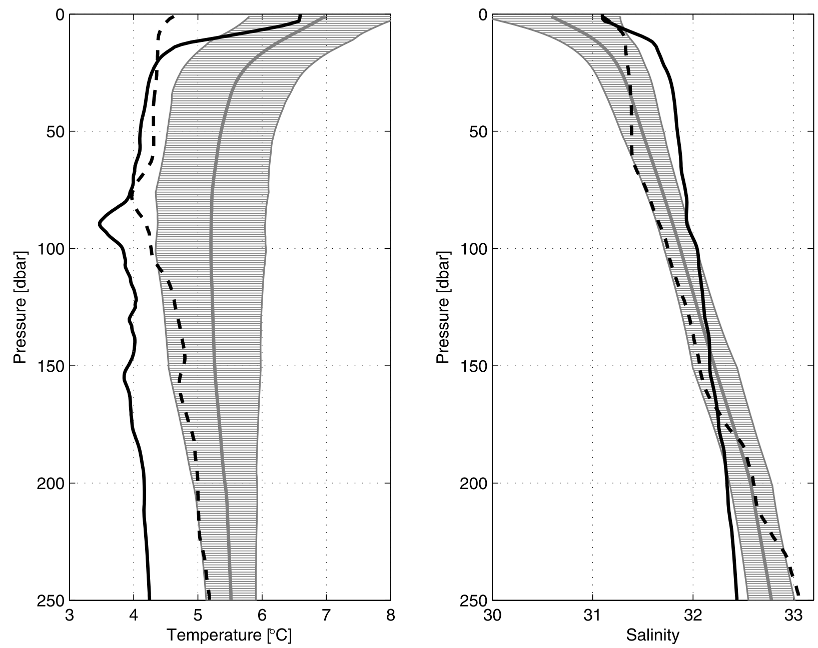

Figure 3. Mean May (left) temperature (°C) and (right) salinity profiles (gray line) ± one standard deviation (gray shading) based on 41 May CTD profiles from 1970 to 2008. The solid black profiles are for 8 May 2007 and dashed profiles are for 4 May 2008.

being easterly. We use the meteorological convention to indicate wind direction throughout the paper.

[7] Monthly coastal freshwater discharge anomalies from the Alaska-British Columbia boundary to 150°W are estimated from *Royer*'s [1982] hydrological model. These are significantly correlated with upper ocean salinities at GAK1 [*Royer*, 2005; *Weingartner et al.*, 2005] and the baroclinic transport in the ACC [*Weingartner et al.*, 2005].

3. Results and Analysis

3.1. Hydrographic Observations at GAK1 in Spring 2007 and 2008

[8] GAK1 measurements in May 2007 indicated anomalous vertical profiles of temperature and salinity in comparison to vertical profiles of their means and standard deviations (Figure 3) for May based on the 1970–2008 record (41 profiles). Although the upper 10 m of the water column had already warmed to ~7°C in early May 2007, temperatures below 10 m were nearly uniform at ~4°C and 1.3–1.4°C below average. Overall, May 2007 temperatures were ~2σ (σ, standard deviation) lower than the mean; they were ~1.5σ below average from 10 to 100 m and ~3σ below average from 200 to 250 m. However, in May 2008, the largest temperature anomalies were at the surface (−2σ) and ~2.5°C lower than average. Temperature anomalies were −2°C (−1.8σ) at 10 m and −1°C (~−1σ) at 100 m. At depths greater than 100 m, temperatures were ~0.5°C below average and thus within the average May range of variability.

[9] In May 2007, salinity varied from 31.6 at 10 m to 32.4 at 250 m, with the salinity anomalies decreasing nearly linearly from +0.6 (1.5 σ) at 10 m to −0.3 (−1.5σ) at 250 m. This anomalous salinity distribution resulted in a relatively small density difference (~0.4 kg m^{-3}) between the upper (0–100 m) and lower (100–250 m) layers, which we argue induced the anomalously low spring temperatures throughout the water column. In contrast, salinities in May 2008 were within 1σ of the average and varied from 31.3 at 10 m to 33.0 at 250 m. In particular, salinities were 0.1–0.5 above average near the surface (0–30 m), 0.1–0.2 lower than average between 30 and 170 m, and nearly 0.3 above average at 250 m. Thus, the (salinity-controlled) density difference between the upper and lower layer was 0.7 kg m^{-3} and near the May average of 0.65 kg m^{-3}. The winters and springs of 2007 and 2008 constitute the longest continuous period since the early 1970s of anomalously low temperatures observed on the GOA shelf (Figure 4).

[10] To place the GAK1 observations in a cross-shelf context, we examined May temperature and salinity anomalies constructed for the 1998–2008 period from individual station means along the Seward Line. Figures 5 and 6 show average upper- (0–100 m) and lower-layer (100–250 m) temperature and salinity anomalies between the inner shelf (GAK1) and the shelf break (GAK9). Stations GAK10–13 are not shown because these are located on the slope, where conditions may be influenced by basin processes. May temperature anomalies are generally consistent across the shelf and were positive (>0.5°C) in 1998, 2000, and 2003. Large negative (0–100 m) temperature anomalies (~−0.5 to −1°C) occurred in May 2002, 2007, and 2008 and extended across the entire shelf, although the largest anomalies (<−1°C) occurred on the inner shelf in 2007 and 2008. Lower-layer (100–250 m) temperatures were above average in 1998 and 2003 (>0.5°C) and below average in 2007 and 2008, with the largest negative anomalies of < −1°C across GAK1–5 during May 2007. The cross-shelf temperature anomaly pattern distribution in May 2008 was similar, although the anomaly magnitudes were smaller than those of 2007.

[11] Upper-layer salinity anomalies along the Seward Line (Figure 6) were negative in 1998, generally positive from

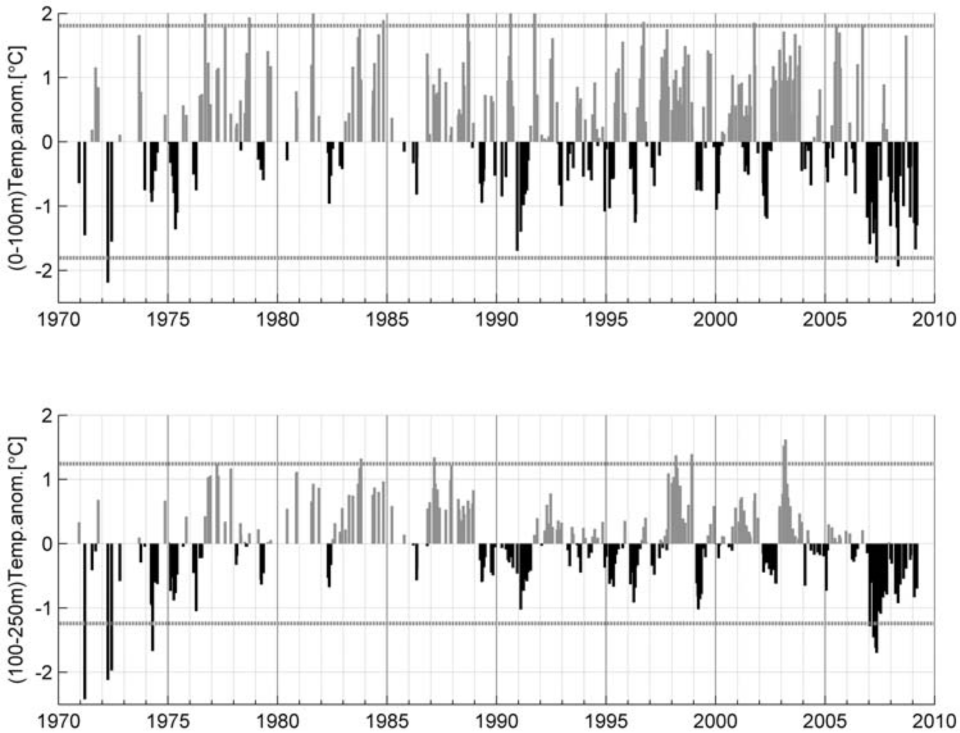

Figure 4. Detrended (top) upper- (0–100 m) and (bottom) lower-layer (100–250 m) GAK1 monthly temperature anomalies (°C) computed from CTD profiles acquired ~1970 to January 2009. The dotted lines indicate ±2 standard deviations. Updated from *Royer and Grosch* [2006].

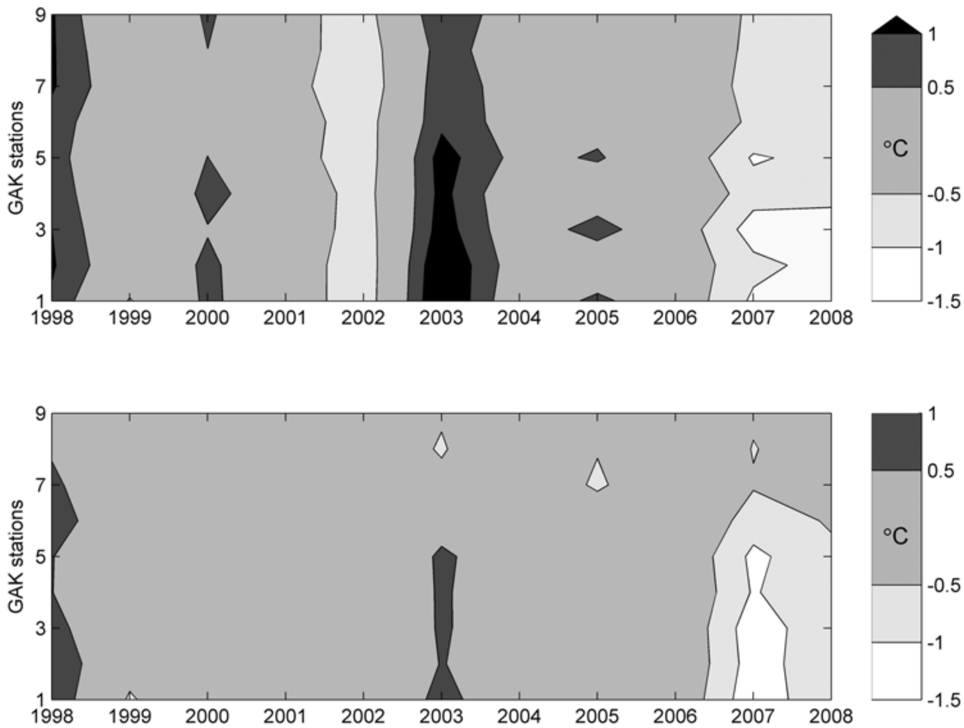

Figure 5. (top) Upper- (0–100 m) and (bottom) lower-layer (100–250 m) Seward Line May temperature anomalies (°C) between the coast (GAK1) and the shelf break (GAK9). Anomalies are computed for the period 1998–2008.

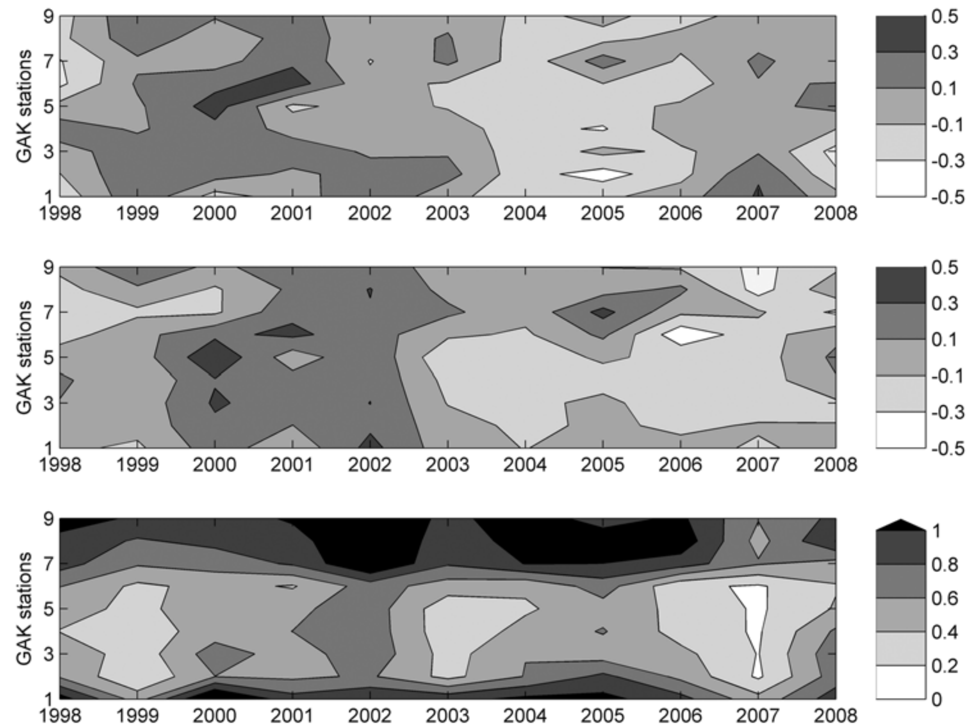

Figure 6. (top) Upper- (0–100 m) and (middle) lower-layer (100–250 m) Seward Line May salinity anomalies between the coast (GAK1) and the shelf break (GAK9), and (bottom) the salinity difference between upper and lower layers.

1999 to 2002, and negative again from 2003 to 2006. Upper-layer salinities in 2007 and 2008 were near average on the middle and outer shelf (GAK4–9) but above average in 2007 and below average in 2008 on the inner shelf (GAK1–4). The lower-layer salinity anomalies follow the upper-layer trend, except in 2007, when lower-layer salinities were fresher than average across the shelf. In May 2007, salinity differences between the upper (0–100 m) and lower (100–250 m) layers were the smallest (<0.2) observed across GAK2–6 (again indicating anomalously weak stratification) and coincided with the largest negative temperature anomalies in the lower layer. The outer-shelf salt stratification (GAK7–9) in 2007 was also the weakest observed among all these years.

3.2. Onset of the Cooling in November 2006

[12] In this section we use the nine years (1998–2008) of moored GAK1 temperature and salinity to show the annual cycles of 30–100 m averaged temperatures (Figure 7) and stratification (Figure 8; based on the vertical density difference between 30 and 100 m) to highlight November 2006 as the beginning of the recent cooling. The monthly mean temperature graph is nearly sinusoidal with a maximum in September and October (~9.5°C) and a minimum of ~4.5°C in March (Figure 7) when the stratification is also weakest (Figure 8).

[13] In 2006, the mean temperatures at 30–100 m were near average in July, but they were ~1°C above average from August through mid-October. November temperatures then declined by more than 3°C, from ~9.5°C at the beginning of the month to 6°C by the end of the month, with this decrease being nearly twice the mean November cooling rate. Moreover, this precipitous temperature decline signaled the transition from a relatively warm early fall to the coolest winter/spring in the GAK1 mooring record. Temperatures continued to decrease rapidly through December 2006 before falling more slowly to a minimum of ~3°C by late March 2007. Nevertheless the January–March temperature decrease of 2.5°C outpaced the mean temperature decrease of ~1.5°C. Temperatures then slowly increased to ~4°C in early May, but the mean temperatures at 30–100 m remained below average from May through September 2007. Temperatures did return to average by October and remained so through late January 2008. Upper-layer waters then cooled more rapidly than average with abnormally low temperatures (<4°C) reached in March 2008. Note that the most rapid temperature decrease (Figure 7) and erosion in stratification (Figure 8) occurred in November 2006, so a more detailed investigation into atmospheric and oceanographic conditions at this time is warranted.

[14] Figure 9 shows the moored GAK1 temperature and salinity record from November 2006 along with nearby QuikSCAT winds and NCEP air temperatures. Oceanographic conditions in November evolved in response to two strong cooling events associated with westerly and northerly winds, as well as adjustments deep in the water column associated with up- and downwelling winds. The first strong cooling event began around November 7, when strong westerly (>10 m s^{-1}) winds, with air temperatures of between −1°C and −5°C, blew for several days. The shelf response included upwelling reflected by a sudden decrease (increase) in near-bottom temperature (salinity) and surface cooling. A subsequent shift to strong (15 m s^{-1}) easterly winds and an

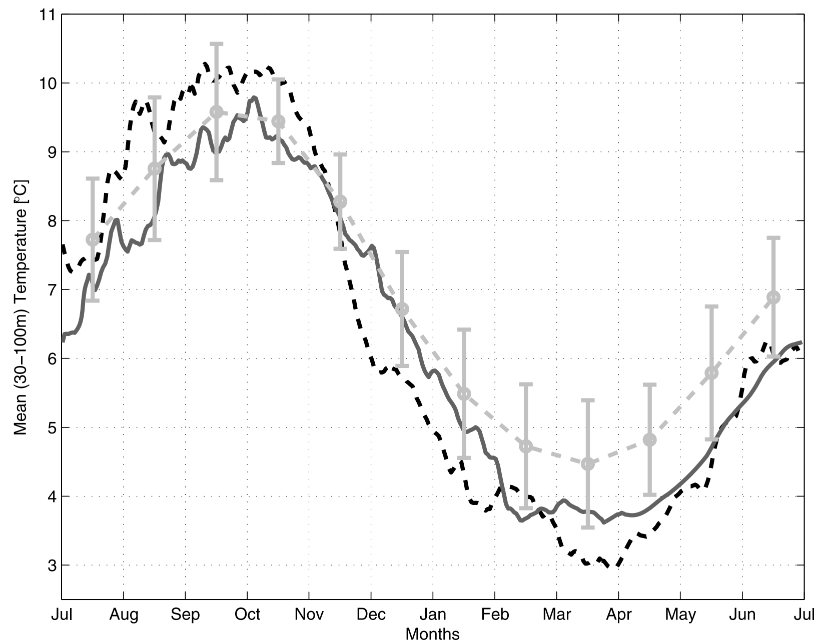

Figure 7. Seasonal cycle of mean 30–100 m GAK1 temperatures (°C) averaged from moored temperature recordings at 30, 60, and 100 m. Temperatures from 2006–2007 (black dashed curve) and 2007–2008 (solid curve) are compared with monthly means from the mooring records from 1998 to 2008 (shaded dashed curve with circles). Vertical bars on monthly means indicate one standard deviation. Monthly means are plotted in the middle of the month, and cross-ticks indicate the first day of each month.

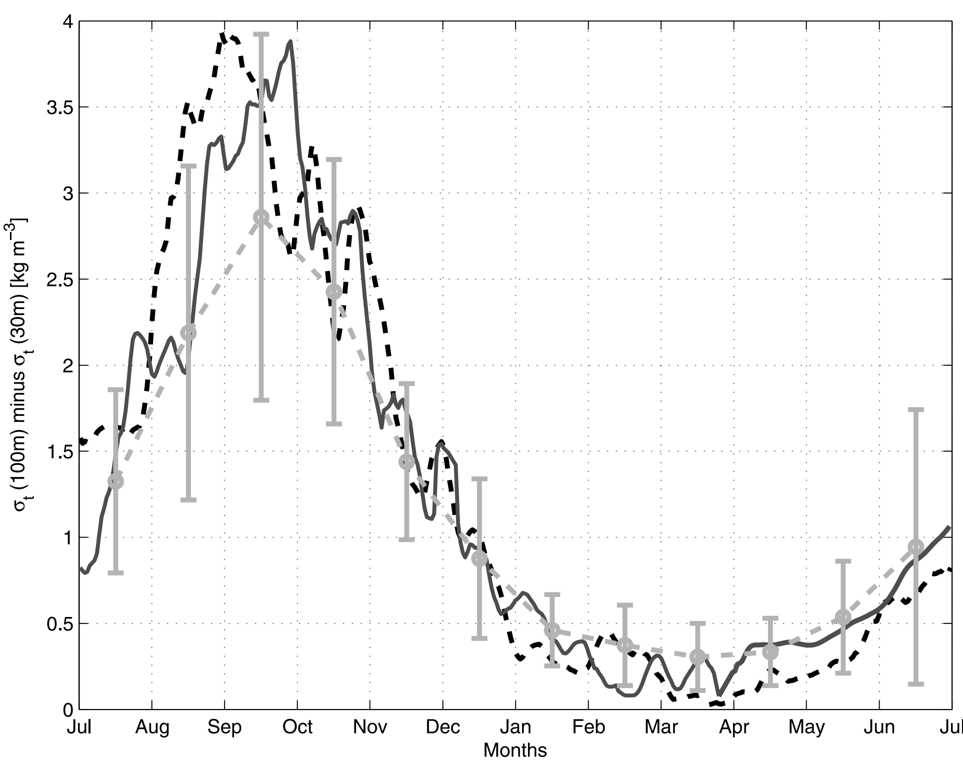

Figure 8. Seasonal cycle of the σ_t difference (kg m^{-3}) between 100 and 30 m as a measure of stratification computed from moored temperature and salinity recordings at 100 and 30 m. σ_t-differences from 2006–2007 (black dashed curve) and 2007–2008 (solid curve) are compared with monthly means from the mooring records from 1998 to 2008 (shaded dashed curve with circles). Vertical bars on monthly means indicate one standard deviation. Monthly means are plotted in the middle of the month, and cross-ticks indicate the first day of each month.

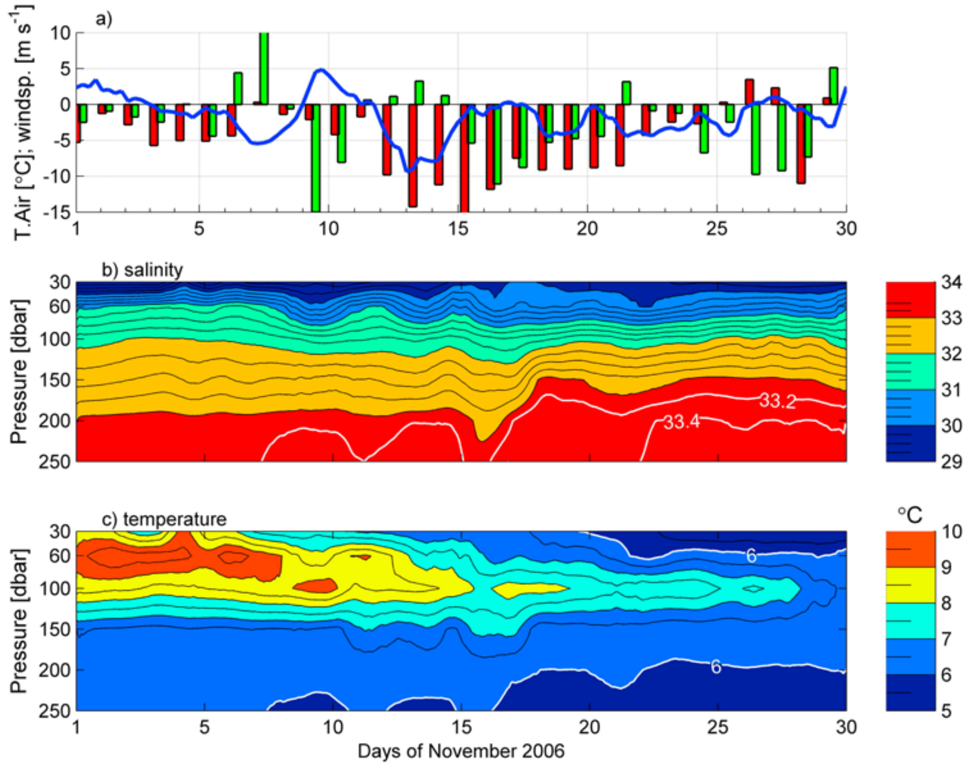

Figure 9. Atmospheric and hydrographic observations from November 2006. (a) NCEP 4 times daily air temperatures (°C, blue line) at 60°N, ~149°W, meridional (red bars) and zonal (green) QuikSCAT wind components (m s^{-1}) (positive indicates southerly and westerly); GAK1 moored (b) salinity; (c) temperature (°C), at 30, 60, 100, 150, 200, and 250 m. Temperature and salinity are smoothed using a 1 day moving average.

increase in air temperatures to 5°C returned bottom thermohaline properties to those of early November and surface cooling rates diminished. Beginning on 12 November, two strong northerly and northeasterly wind events of (10–15 m s^{-1}) advected cold air (~−10–0°C) over the shelf, which accelerated surface cooling. These (downwelling-favorable) winds caused a depression in the isohalines below 200 m of ~25 m and decreased salinities to < 33.2 at 250 m. Winds then veered northerly and the 33.2 isohaline rebounded to between 160 and 200 m depth, where it remained until December. These northerlies involved the highest wind speeds observed in the mooring record and they also coincided with the second-lowest November monthly mean air temperatures archived in the (1948–2008) NCEP record.

[15] The increase in deep salinities was also accompanied by a decrease in deepwater temperatures to < 6°C between 200 and 250 m depth. The source of this cool, salty water is most likely the mid- and outer shelf, ~100 km seaward of GAK1 (based on data from fall and winter cruises collected during the NEP GLOBEC program, available at http://www.ims.uaf.edu/GLOBEC/results). Since the mean GAK1 November/December salinity at 200 m is 32.7 ± 0.3, a November salinity of 33.2 at depths less than 200 m is rare; indeed it has been observed only twice among the 24 November/December GAK1 CTD profiles in the archive. These anomalously high salinities persisted through March 2007 so that deep freshening, which normally occurs in late winter, occurred much later than normal, as shown in the next section.

[16] We next show that freshwater runoff was also anomalously low in November 2006, so the combination of reduced runoff and the deep salinity influx weakened the stratification early in fall. The diminished stratification and the strong cooling may have preconditioned the water column in late fall for the deep cooling that developed subsequently.

3.3. Evolution of the 2006–2007 Cooling Compared to Other Years

[17] We now compare the 2006–2007 cooling with the anomalously cold winters of 2001–2002 and 2007–2008 and the relatively warm winter of 2000–2001. (The other years in the mooring record were similar to 2000–2001 and are not discussed). Our comparison is based on the July to June temperature and salinity time series for these four years (Figures 10a–10d). Companion plots of monthly anomalies of freshwater runoff, NCEP air-sea heat fluxes (where negative anomalies imply increased heat loss to the atmosphere), and the meridional and zonal QuikSCAT wind anomalies are shown in Figure 11. The monthly anomalies, based on the 2000–2008 period, are normalized by their maximum absolute value.

[18] In the months following November 2006, the cold surface signal deepened (Figure 10a), due to wind mixing and cooling and anomalously low runoff that persisted through

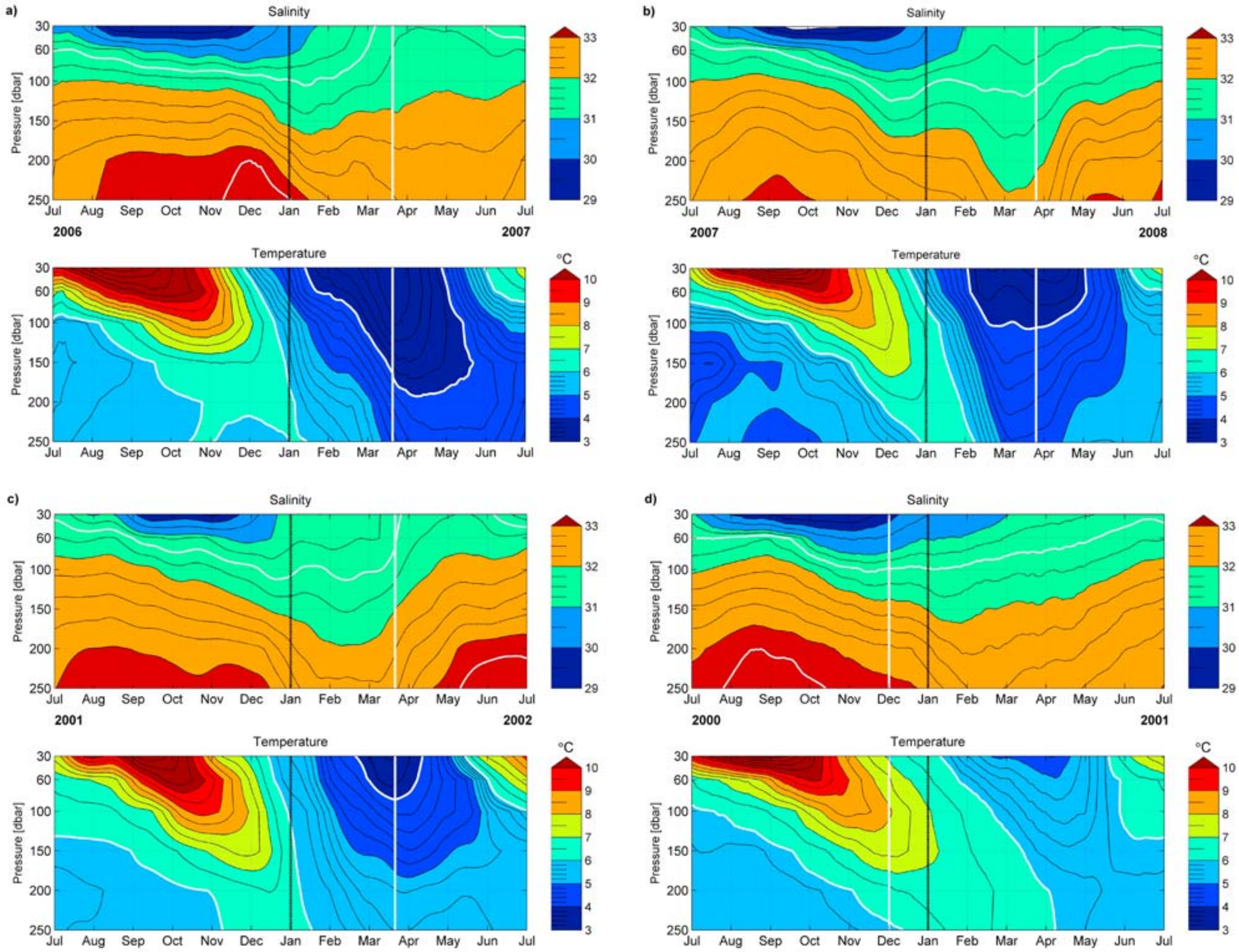

Figure 10. Salinity and temperature (°C) from the GAK1 mooring recorded at 30, 60, 100, 150, 200, and 250 m, shown from 1 July to 30 June of (a) 2006–2007, (b) 2007–2008, (c) 2001–2002, and (d) 2000–2001. Black dashed lines mark 1 January as a reference, and white vertical lines mark the data gap between mooring recovery and deployment. Isotherms of 4°C and 6°C and 31.5 and 33.25 isohalines are highlighted by white contours. Note that isohaline increments vary between 0.25 and 0.5, and isotherms between 0.2°C and 0.5°C as indicated in color legends. Data are smoothed using a 7 day moving average.

winter 2007. Hence, by mid-January, temperatures in the upper 60 m decreased to < 4°C (Figure 10a); these temperatures were attained a month or more earlier than the other cold winters of 2001–2002 and 2007–2008. The low winter runoff also led to the very weak stratification of March 2007 (which together with 1971, 1996, and 1999 rank among the weakest stratifications observed in March in the GAK1 record). The weak stratification and the large air-sea heat losses associated with strong northerly winds in March 2007 (Figure 11) resulted in water temperature of < 4°C at 200 m and lowered temperatures between 30 and 100 m to <3°C for about 1 week near month's end.

[19] March 2007 ended with the development of anomalously strong downwelling winds that persisted through May (Figure 11). These winds likely contributed to the persistence of anomalously cold water at depth through the summer and fall of 2007 (Figure 10b) due to enhanced deep mixing and possibly delayed the subsurface, onshore flow of more saline and warmer outer-shelf water that occurs each spring [*Royer*, 1975; *Xiong and Royer*, 1984; *Weingartner et al.*, 2005]. That delay could have contributed to the fresh bottom salinity anomaly observed in May 2007 (Figure 3). The subsurface negative temperature anomalies formed in spring 2007 waned by late December 2007, when bottom temperatures reached ~6°C (Figure 10b). While the timing of the arrival of 6°C water near the bottom (250 m) and its duration can vary from year to year, it usually occurs in late fall due to increased coastal downwelling and is responsible for the late fall/early winter deep (>150 m) temperature maximum on the northern GOA shelf [*Xiong and Royer*, 1984; *Royer*, 2005; *Weingartner et al.*, 2005] as evident in the four years shown (Figures 10a–10d).

[20] In contrast to 2006–2007, the anomalously low winter and spring temperatures of 2007–2008 (Figure 10b) and 2001–2002 (Figure 10c) were confined to the upper (0–100 m) layer. Furthermore, November 2007 had anomalously small air-sea heat fluxes and large freshwater runoff (>50% above the November average; see Figure 11). Thereafter, discharge anomalies were slightly negative but the stratification did not appreciably decrease until February. Anomalously

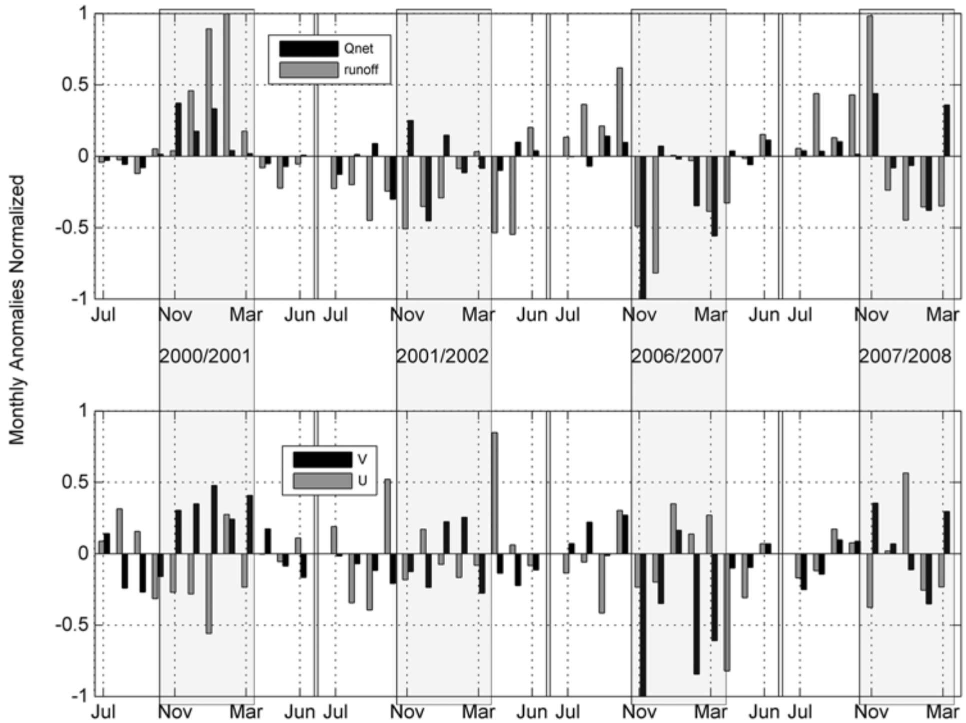

Figure 11. Monthly normalized anomalies (2000–2008): (top) NCEP net heat flux (W m^{-2}) (black bars); freshwater runoff (km^3 s^{-1}) (gray bars); (bottom) meridional (black) and zonal (gray) QuikSCAT components (m s^{-1}), where negative anomalies indicate winds from the north and from the east. Shading highlights the cooling season from November to March. Months shown correspond to the mooring time series in Figures 10a–10d.

strong air-sea heat fluxes only developed in late January and February 2008 (Figure 11), so that cooling to temperature <4°C was confined to between the surface and 100 m (Figure 10b) because of the stratification induced by large freshwater discharge rates in fall 2007. The high fall discharge coincided with anomalous southerly winds, which enhanced precipitation and runoff in the northern GOA [*Weingartner et al.*, 2005].

[21] Contrary to the large coastal runoff rates in fall 2007, below-average runoff from July 2001 to January 2002 (Figure 11) resulted in elevated surface salinities and reduced upper-layer stratification (Figure 10c). Oceanic winter heat loss was anomalously strong in October and December 2001 (Figure 11) and cooled the upper 100 m to <5°C by late January 2002 (Figure 10c). Runoff remained below average from January through April (except March), but air-sea heat flux anomalies were both positive and negative from November through April, so that in spite of the reduced stratification, water temperatures of < 4°C were confined to the upper ~60–80 m (Figure 10c).

[22] The fall and winter of 2000–2001 provides a contrasting example from the cold years. Then ~6°C water persisted from December through April at 250 m (Figure 10d), and large freshwater runoff rates and anomalous southerly winds prevailed throughout the winter (Figure 11) so that relatively high salinities persisted at the bottom and low salinities near the surface. In aggregate, these conditions enhanced stratification and impeded vertical mixing. In addition, heat fluxes were anomalously positive during the winter of 2000–2001.

[23] In summary, the mooring record, although of limited duration, suggests that both winter cooling and reduced runoff play synergistic roles in promoting deep (>100 m) cooling. Runoff structures the shelf stratification, while the air-sea heat flux extracts heat from the ocean. Consequently, we hypothesize that anomalously low temperatures at depth develop in winters of anomalously large air-sea heat fluxes and low fall/winter runoff. We next test this hypothesis by examining atmospheric forcing and thermal anomalies over the entire GAK1 hydrographic record from ~1970 to the present.

3.4. Atmospheric Forcing Parameters and Their Interannual Variability at GAK1

[24] Our analysis is based on winter (November–March) anomalies (computed for the period from 1970 to the present, i.e., over the length of the GAK1 record) of atmospheric parameters (Figure 12). We use salinity measurements from the monthly profiles to approximate upper-layer stratification as the vertical salinity difference between 20 and 100 m. We consider spring temperature anomalies in the upper (0–100 m) and lower (100–250 m) layers. To assess the integrated oceanic response to wintertime forcing, we averaged March–May temperature profiles and compare these means with winter-averaged anomalies of air temperatures, freshwater runoff, meridional and zonal winds, and latent and sensible heat fluxes. Changes in ocean temperatures between March and May are mainly limited to the upper ~10 m, and this averaging approach increases the yearly data density,

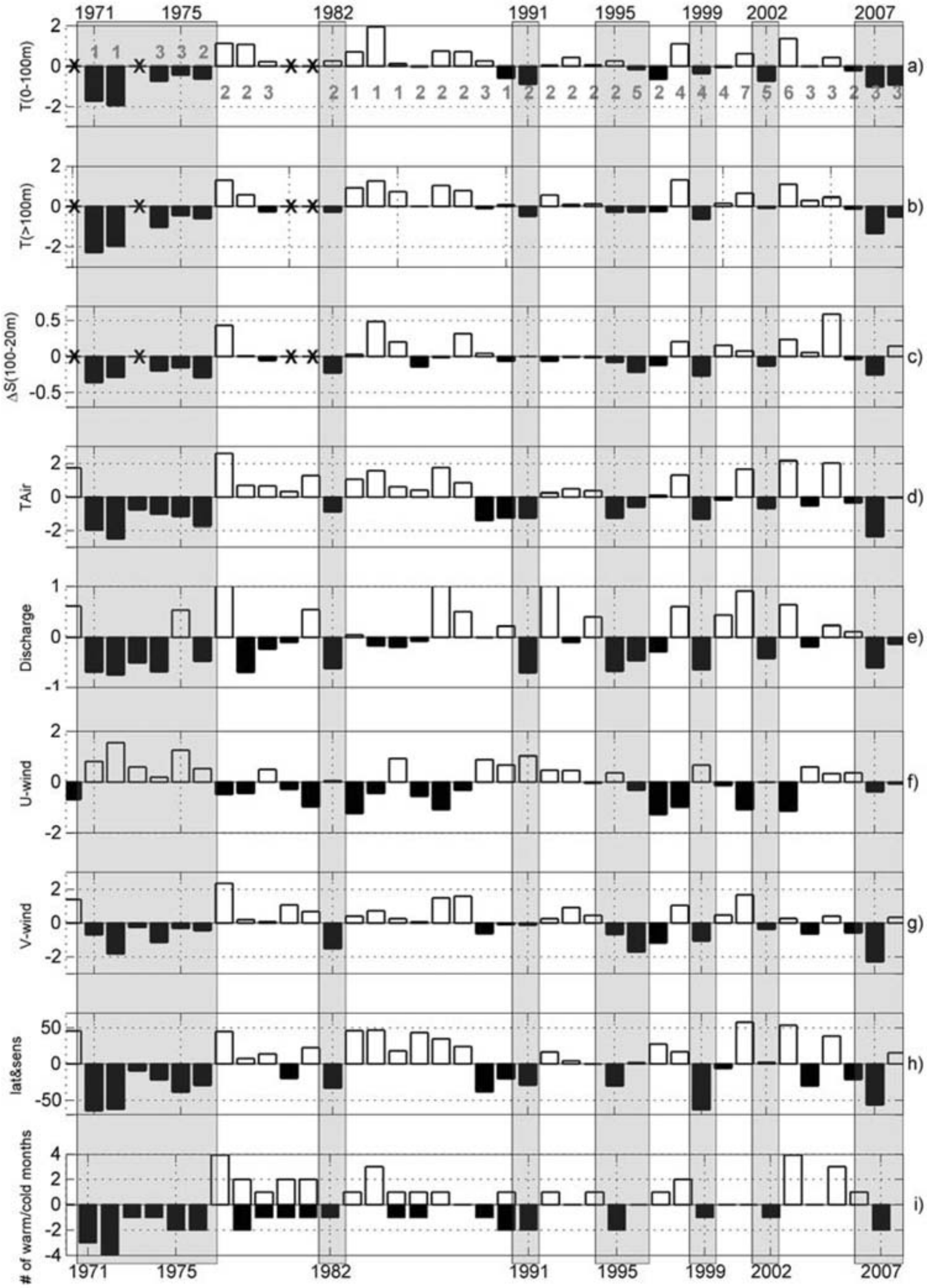

Figure 12. (a) Upper-layer (0–100 m) and (b) lower-layer (100–250 m) temperature anomalies (°C), and (c) the salinity difference between 100 and 20 m, all averaged from February–May CTD profiles, compared with winter (November–March) anomalies from 1970 to the present of (d) NCEP air temperatures (°C), (e) freshwater runoff (km^3 s^{-1}), (f) zonal and (g) meridional wind speed (m s^{-1}), (h) latent and sensible heat flux (W m^{-2}); and (i) the number of months with anomalously high or low air temperatures between November and March. Gray numbers in Figure 12a indicate the number of CTD casts used for averaging. Shaded boxes highlight years with anomalously low (<1σ) deep (100–250 m) temperatures. Solid bars indicate negative anomalies. Note that there were no data in 1970, 1973, 1980, or 1981 for Figures 12a–12c.

since sampling coverage was relatively sparse prior to the late 1980s (Figure 2).

[25] Large negative (>−1°C) temperature anomalies in the surface layer (0–100 m) occurred in the early 1970s and in 2007–2008, whereas weaker negative anomalies (~−0.5°C) occurred in 1982, 1991, 1995, 1996, 1999, and 2002. Each of these years had anomalously low air temperatures and all years (except 1975) had anomalously low winter runoff. Overall, 2006–2007 had the largest anomalies in meridional winds, air temperatures, and coastal runoff, with similar anomalies occurring in 1971–1972 and 1972–1973, when the lowest water temperatures were observed. By contrast, 1977,

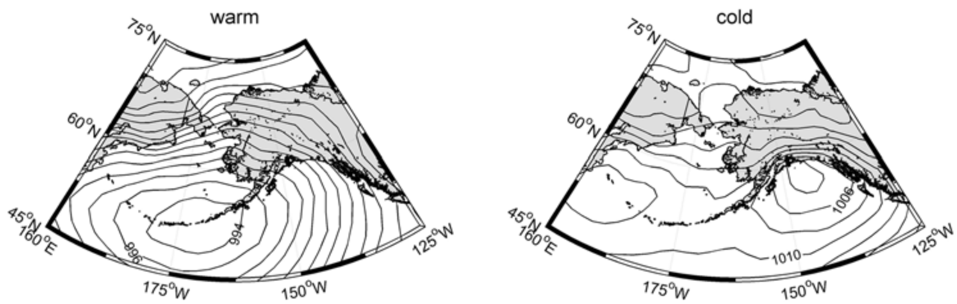

Figure 13. Averaged SLP (mbar) distribution during months of anomalously >|1σ| (left) high and (right) low November–March air temperatures at GAK1. Contours are in 2 mbar increments.

1984, 1998, and 2003 were anomalously warm (0–100 m temperature anomalies exceeded +1°C) and coincided with positive anomalies in air temperature, runoff, stratification, and positive heat loss anomalies. The statistical relationship among temperature, runoff, and heat fluxes is discussed in more detail below.

[26] During cold winters, precipitation on land accumulates as snow, so anomalously low runoff and air temperatures tend to be correlated. Indeed, this is factored into *Royer*'s [1982] model of coastal freshwater discharge. However, upper ocean salinity and vertical salinity gradients between 20 and 100 m at GAK1 both reflect the freshwater discharge [*Royer*, 1982; *Weingartner et al.*, 2005], and therefore provide independent measures of Royer's model. With the exception of 1975, years with anomalously low, deep (100–250 m) temperatures coincide with below-average winter runoff and weak salinity gradients between 20 and 100 m (Figure 12). In fact, stepwise regression (n = 35) shows that ~81% of the variance in lower-layer (100–250 m) temperature anomalies are explained by salinity stratification and air temperatures (all regressions are significant at the 99% confidence level). Variations in air temperatures and in the vertical salinity gradient alone explain 75% and 67% of the deepwater temperature variance, respectively. Meridional winds explain 57%, and heat fluxes and runoff explain 60% and 42% of the variance, respectively, although both heat fluxes and runoff depend (in part) on air temperature.

[27] We next investigated the surface atmospheric pressure distribution over the North Pacific during anomalously cold and warm winter months during the ~40 years of the GAK1 time series. We identified months with anomalously low (> |1σ|, where σ ranges from 1.4°C to 2.3°C; 33 months) and high (32 months) GAK1 NCEP air temperatures between November and March from 1970 to 2008 (Figure 12i) and then constructed average sea level pressure (SLP) patterns based on this sorting (Figure 13). The warm pattern is characterized by an extensive AL over the central Aleutians, which results in southerly winds and large heat and moisture transport to the northern GOA. In contrast, the cold pattern consists of two weak low-pressure cells, one over the western Aleutians and the other centered in the northern GOA, and a high-pressure ridge over the mainland. This pattern implies northeasterly winds that advect cold air over the western GOA shelf (Figure 11) and also favors seaward ageostrophic winds that are channeled through gaps in the coastal mountains and bays along the south coast of Alaska [*Macklin et al.*, 1988]. Consequently, the (November–March) meridional wind component near GAK1 is significantly correlated with NCEP GAK1 air temperatures (r = 0.74, p < 0.01) and with coastal runoff (r = 0.53, p < 0.01).

[28] In warm winters, the SLP gradient between GAK1 and the central Aleutians is large, whereas it is small in cold years. We found that the maximum correlation between GAK1 air temperatures and the SLP difference was between NCEP grid points at 60°N, 150°W (nearest to GAK1) and 54°N, 176°W for the January–March period (r = 0.84, p < 0.05). Hence, a significant fraction (~70%) of the winter air temperature variability is explained by this SLP gradient, which is a function of the location of the AL.

[29] This finding agrees with that of *Rodionov et al.* [2007], who investigated atmospheric conditions in the Bering Sea in relation to the AL and found only weak or no correlations (depending on the method) between Bering Sea air temperatures and the North Pacific Index (NPI) [*Trenberth and Hurrell*, 1994], a measure of AL strength. Similarly, we found no correlations between GAK1 winter air temperatures and the NPI, which underlines the importance of the location of the AL rather than its strength on the northern GOA climate.

3.5. On the Timing of Cooling Events

[30] From October to March, northern GOA waters lose heat to the atmosphere on average, while coastal freshwater discharge is at its annual minimum in February and March [*Royer*, 1982]. We suggest that cooling events and discharge anomalies during fall and late winter (e.g., October–November and March–April) may have a comparatively greater effect on stratification and the vertical temperature distribution than midwinter events. For example, coastal freshwater discharge is a maximum in fall (mean November runoff is ~40×10^3 m^3 s^{-1}) so that a 30% reduction in fall runoff (as occurred in November 2006) affects the shelf freshwater reservoir more than a similar reduction from January through March (when runoff is ~10–15 × 10^3 m^3 s^{-1}). Since the flushing time scale for the northern shelf is ~8 months [*Weingartner et al.*, 2005], the memory of fall runoff anomalies is retained through winter. Moreover, runoff is positively correlated with the along-shelf transport; hence, a negative runoff anomaly in fall will also reduce the along-shelf transport of heat [*Weingartner et al.*, 2005]. Similarly, large negative air-sea heat flux anomalies during the transition from cooling to warming in late March may extend the cooling season and, more importantly, may allow wind mixing and cooling to extend

much deeper into the water column since stratification is at its annual minimum in late winter (Figure 8).

[31] Consequently, we hypothesize that anomalously low temperatures at depths > 100 m are a consequence of below-average fall runoff and above-average late-winter cooling. The early 1970s and 1991, 1995, and 2007 each had anomalously low late-winter/spring temperatures (Figure 12) and in each of these years late-fall or early winter runoff was below average. Coastal runoff and heat flux anomalies in November 2006 and March 2007 were among the largest negative anomalies recorded, and these coincide with some of the most anomalous hydrographic conditions in the GAK1 record. However, 2006–2007 is the only such winter in which highly resolved time series allowed us to observe the evolution of deep cooling. Hence, while our hypothesis seems plausible, definitive proof awaits further investigation.

3.6. Effect of Downwelling

[32] Recently, *Shcherbina and Gawarkiewicz* [2008, hereafter SG08] examined the relative importance of wind-driven buoyancy flux (WDBF) due to onshore surface Ekman transport versus the buoyancy flux due to atmospheric cooling in the narrow (~10 km), shallow (~50 m) Outer Cape Cod Coastal Current (OCCC). They found that the winter buoyancy loss in the OCCC due to onshore advection of saline surface waters under downwelling winds exceeded the buoyancy loss due to air-sea heat exchange so that WDBF enhanced deep mixing in the OCCC. We made similar estimates of these fluxes for the ACC from the October–March monthly NCEP net heat fluxes, alongshore wind stresses, and the surface density gradient between GAK1 and GAK4 estimated from Seward Line transects. On the northern GOA shelf, density increases offshore (to the south) and the surface density gradients weaken from ~−6 × 10^{-5} kg m^{-4} in October to ~−2 × 10^{-5} kg m^{-4} in March. For a mean winter alongshore wind velocity of −4 m s^{-1} [*Weingartner*, 2007], the WDBF (= $-[\tau^x/f\rho_0]g\rho_0^{-1}\partial\rho/\partial y$, where ρ is the density, ρ_0 is a reference density, y is the cross-shore coordinate, τ^x is the alongshore wind stress, f is the Coriolis parameter, and g is gravity) varies from ~−2 × 10^{-7} m^2 s^{-3} in December to ~−6 × 10^{-8} m^2 s^{-3} in March (i.e., shoreward buoyancy flux in December is about three times greater than in March). Buoyancy losses due to atmospheric cooling are smaller and vary seasonally from ~−2 × 10^{-8} to −5 × 10^{-8} m^2 s^{-3}. Our estimates of the WDBF and the atmosphere-ocean buoyancy flux for the ACC are similar in magnitude to those of the OCCC per SG08.

[33] One might expect that the destabilizing influence associated with shoreward buoyancy flux would enhance deep mixing of winter-cooled surface waters in the ACC and GOA shelf. This does not appear to be the case, however, since winter downwelling anomalies are anticorrelated (r = −0.3, $p < 0.1$) with the vertical salinity gradient and water temperatures (r = ~−0.57, $p < 0.01$, for both the 0–100 m and the 100–250 m portions of the water column). The reasons for this are likely several. First, the mean sea level pressure maps for warm and cold winters (Figure 13) suggest that winters with strong downwelling are associated with above-average air temperatures and greater-than-average runoff. Indeed, winter downwelling winds are anticorrelated with winter air temperatures (r = −0.64, $p < 0.01$), net air-sea heat fluxes (r = −0.72, $p < 0.01$), and runoff (r = −0.39, $p < 0.1$). Second, downwelling-favorable winds not only drive a small onshore heat flux (which we estimate to be ~25 W m^{-2} on average) but also enhance the alongshore advection of oceanic heat and freshwater over the northern GOA shelf. Finally, *Williams et al.* [2007] and SG08 showed that the cross-shelf transport is sensitive to the structure and strength of the haline fronts that bound the offshore edge of the ACC and OCCC. The strength of the ACC front in winter is at least partially dependent upon the winter discharge [*Weingartner et al.*, 2005]. In aggregate, our results suggest that the winter evolution of temperatures on the GOA shelf is a consequence of three-dimensional circulation and mixing processes involving a complex interplay among the air-sea heat flux; the ocean heat flux convergences; the stabilizing influence of runoff; and the destabilizing effects of cooling, vertical mixing, and the WDBF. A profitable area for future research would be to examine these influences using simple process models.

4. Summary and Conclusion

[34] Long-term temperature and salinity observations at hydrographic station GAK1 on the northern GOA shelf revealed anomalously low water temperatures in the winters and springs of 2007 and 2008. Indeed in spring 2007, temperatures throughout the water column were the lowest observed in ~35 years (Figure 4) and the vertical stratification was also unusually weak. Conditions in spring 2008 also included some of the coldest waters in the record, but the temperature anomalies then were confined to the upper 100 m of the water column because of salt stratification below this depth. Our results suggest that winter coastal runoff, by modulating upper-ocean salinities and winter stratification, exerts an important influence on the temperature distribution in the northern GOA. Our analysis of NCEP atmospheric variables, coastal freshwater discharge, and QuikSCAT winds also suggests that the timing of winter cooling events, i.e., events at the beginning (October–November) and end (March–April) of the cooling season, are important in shaping the hydrographic conditions for the following spring and summer.

[35] For example, in November 2006, net air-sea heat fluxes, coastal runoff, and northerly winds were among the most anomalous on record. These conditions cooled the water column and weakened the stratification. Deep shelf processes may also have contributed to the inflow of colder, saltier water following a sudden relaxation of strong downwelling-favorable winds. Cooling continued from December through February although the air-sea heat fluxes were close to their climatological mean, but the stratification remained below average due to below-average winter runoff. Consequently, enhanced cooling and deep mixing of the GOA shelf occurred in March 2007 as a result of unusually severe cold air outbreaks. Interestingly, before vigorous cooling began in November 2006, the stratification and thermal anomalies in September and October 2006 were above average, suggesting that summer and early fall upper-ocean anomalies can be obliterated rapidly on the northern GOA shelf in late fall and winter.

[36] About two thirds of the alongshore baroclinic geostrophic transport on the GOA shelf is within the ACC [*Weingartner et al.*, 2005], which advects freshwater and heat along the GOA coast. In 2006–2007, the alongshore heat and

freshwater advection was likely reduced. While the wind-driven (barotropic) component of the ACC was likely average due to near-average alongshore winds from November to March (Figure 12), the nearshore baroclinic flow (estimated from Seward Line transects) in May 2007 was the lowest among 11 May (1998–2008) transects. This implies that the alongshore heat contribution was diminished and likely contributed to the cooling in 2006–2007.

[37] A comparison of the 2006–2007 cooling season with the relatively cold winters of 2001–2002 and 2007–2008 suggests that preconditioning of the shelf stratification in late fall through cooling and reduced runoff along with late-winter cooling events may be necessary ingredients for the production of deep temperature anomalies in spring. Deep (>100 m) temperature anomalies formed in winter may persist for 6–9 months and thereby be longer lived than upper-layer anomalies, since the deep anomalies are subjected to dynamics governed by the alongshore and cross-shore winds, whereas thermodynamic processes govern the variability in the upper layer. This is supported by the observations from summer through early winter 2007, which showed that the subsurface thermal anomalies established in spring 2007 persisted into the following winter despite the absence of substantial cooling until January–February 2008 (Figures 11 and 12).

[38] Retrospective analyses of the GAK1 time series and the recent cooling events underscore the importance of coastal runoff and upper-ocean salinity on deep (>100 m) shelf temperatures in the northern GOA. The upper-layer (0–100 m) salinity and the lower-layer (100–250 m) temperature anomalies are significantly correlated throughout the GAK1 time series (1970 to the present; $r = -0.33$, $p < 0.01$, $n = 315$), particularly during anomalously cold ($> 2\sigma$) springs (March and April; $r = -0.64$, $p < 0.05$, $n = 12$). Our finding is consistent with *Royer*'s [2005] report of maximum correlations ($r = 0.34$–0.39) between freshwater runoff and GAK1 temperatures at depths greater than 100 m.

[39] Our analysis underscores the importance of freshwater on stratification with ramifications for winter vertical mixing and temperature distribution. While much of the variability in winter heat content is associated with air-sea heat fluxes, the vertical temperature distribution on the GOA shelf is a consequence of three-dimensional circulation and mixing processes involving a complex interplay among air-sea heat fluxes; ocean heat flux convergences; the stabilizing influence of runoff; and the destabilizing effects of cooling, vertical mixing, and the WDBF due to alongshore winds. The aggregate effect of these processes is poorly understood. Moreover, the correct implementation of line or distributed runoff sources in numerical circulation models of the Gulf of Alaska remains problematic [*Dobbins et al.*, 2009]. Until this hurdle is overcome, model predictions of the vertical distribution of springtime temperatures and stratification will remain difficult.

[40] We have argued that the winter/spring temperature distribution in the coastal GOA depends upon fall and winter coastal freshwater discharge. The discharge depends on air temperatures, however, which affect both the moisture content (and, hence, precipitation rates) of the atmosphere via the Clausius-Clapeyron relationship and the partitioning of the precipitation between the liquid (rain) and solid (snow) phases. This affects winter stratification because rain results in relatively rapid runoff, whereas the snow is stored in the surrounding mountains until it melts in the following spring and summer [*Royer*, 1982]. At present, average winter air temperatures around the coast hover near the freezing point [*Brower et al.*, 1988], so slight changes in air temperature can have important consequences on the thermal properties of this shelf. *Intergovernmental Panel on Climate Change* [2007] climate predictions for the GOA region point to warmer and wetter winters, which would result in stronger winter stratification. Our analysis suggests that this will lead to a reduction in the interannual variability of deep winter shelf temperatures, but perhaps to an increase in variability of upper-ocean temperatures since winter mixing will be confined to a shallower depth. Conceivably these changes could include years in which shallow surface layers are quite cold.

[41] The implications of deep cooling and mixing for biological production may be substantial. Weak stratification during spring may result in enhanced surface nutrient concentrations due to deep mixing but may delay the spring bloom. A delay in the bloom and cooler temperatures may affect zooplankton growth due to changes in the phasing of primary production and in zooplankton metabolic rates. It appears that both primary and secondary production was delayed along the Seward Line in May 2007 and 2008 (R.R. Hopcroft, personal communication, 2009). Whether or not 2007–2008 data indicate interannual variability or a transition into a cooler period remains to be seen. However, preliminary analyses suggest that the spring of 2009 was also unusually cold (unpublished data). We note, however, that the 1976–1977 regime shift to warmer ocean temperatures was accompanied by a transition in commercially important fish from crab and shrimp to pelagic fish populations [*Anderson and Piatt*, 1999; *Hare and Mantua*, 2000] and, therefore, had considerable ecological and economical impact for the region.

[42] **Acknowledgments.** M.A.J. acknowledges the support of a University of Alaska Fairbanks (UAF) Center for Global Change Student Award funded by the International Arctic Research Center (IARC) through cooperative agreement ARC 0327664 with the National Science Foundation and a graduate student award from the North Pacific Research Board (NPRB). We thank the Exxon Valdez Oil Spill Trustees Council (EVOSTC; project 070340) for continued support of the GAK1 mooring and monthly CTD data. EVOSTC also supported T.J.W. and S.L.D. Hydrographic sampling along the Seward Line from 1998 to 2004 was carried out as part of NEP-GLOBEC, and since 2005 has been under support of the NPRB (projects 520, 603, 708, and 804). This manuscript is NPRB publication 230. NCEP reanalysis data were provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, from their Web site at http://www.cdc.noaa.gov/. QuikSCAT data are produced by Remote Sensing Systems and sponsored by the NASA Ocean Vector Winds Science Team. Data are available at http://www.remss.com. We thank Steve Okkonen for many helpful discussions, and Andrey Shcherbina and Glen Gawarkiewicz stimulated the discussion in section 3.6. Isaac Schroeder provided advice on the QuikSCAT data. Dave Leech serviced the moorings and collected the monthly CTD casts at GAK1. Russ Hopcroft led CTD data collection efforts along the Seward Line since 2005. We greatly appreciate the comments of the editor and one anonymous reviewer, which substantially improved the manuscript.

References

Anderson, P. J., and J. F. Piatt (1999), Community reorganization in the Gulf of Alaska following ocean climate regime shift, *Mar. Ecol. Prog. Ser.*, *189*, 117–123, doi:10.3354/meps189117.

Brower, W. A., Jr., R. G. Baldwin Jr., C. N. Williams, J. L. Wise, and L. D. Leslie (1988), *Gulf of Alaska*, vol. 1, Natl. Clim. Data Cent., Asheville, N. C.

Childers, A. R., T. E. Whitledge, and D. A. Stockwell (2005), Seasonal and interannual variability in the distribution of nutrients and chlorophyll-a across the Gulf of Alaska shelf: 1998–2000, *Deep Sea Res., Part II*, *52*, 193–216, doi:10.1016/j.dsr2.2004.09.018.

Dobbins, E. L., A. J. Hermann, P. Stabeno, N. A. Bond, and R. C. Steed (2009), Modeled transport of freshwater from a line-source in the coastal Gulf of Alaska, *Deep Sea Res., Part II* , *56*, 2409–2426, doi:10.1016/j.dsr2.2009.02.004.

Hare, S. R., and N. J. Mantua (2000), Empirical evidence for North Pacific regime shifts in 1977 and 1989, *Prog. Oceanogr.*, *47*(2–4), 103–146, doi:10.1016/S0079-6611(00)00033-1.

Intergovernmental Panel on Climate Change (2007), *Climate Change 2007: Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by M. L. Parry et al., pp. 7–22, Cambridge Univ. Press, Cambridge, U. K.

Johnson, W. R., T. C. Royer, and J. L. Luick (1988), On the seasonal variability of the Alaska Coastal Current, *J. Geophys. Res.*, *93*(C10), 12,423–12,437, doi:10.1029/JC093iC10p12423.

Macklin, S. A., G. M. Lackmann, and J. Gray (1988), Offshore-directed winds in the vicinity of Prince William Sound, Alaska, *Mon. Weather Rev.*, *116*, 1289–1301, doi:10.1175/1520-0493(1988)116<1289:ODWITV>2.0.CO;2.

Rodionov, S. N., N. A. Bond, and J. E. Overland (2007), The Aleutian Low, storm tracks, and winter climate variability in the Bering Sea, *Deep Sea Res., Part II*, *54*, 2560–2577, doi:10.1016/j.dsr2.2007.08.002.

Royer, T. C. (1975), Seasonal variations of waters in the northern Gulf of Alaska, *Deep Sea Res.*, *22*, 403–416.

Royer, T. C. (1981), Baroclinic transport in the Gulf of Alaska, part II: Freshwater driven coastal current, *J. Mar. Res.*, *39*, 251–266.

Royer, T. C. (1982), Coastal freshwater discharge in the Northeast Pacific, *J. Geophys. Res.*, *87*(C3), 2017–2021, doi:10.1029/JC087iC03p02017.

Royer, T. C. (2005), Hydrographic responses at a coastal site in the northern Gulf of Alaska to seasonal and interannual forcing, *Deep Sea Res., Part II*, *52*(1–2), 267–288, doi:10.1016/j.dsr2.2004.09.022.

Royer, T. C., and C. E. Grosch (2006), Ocean warming and freshening in the northern Gulf of Alaska, *Geophys. Res. Lett.*, *33*, L16605, doi:10.1029/2006GL026767.

Schroeder, I. (2007), Annual and interannual variability in the wind field and the hydrography along the Seward Line in the northern Gulf of Alaska, Ph.D. dissertation, 80 pp., Old Dominion Univ., Norfolk, Va.

Schumacher, J. D., and R. K. Reed (1980), Coastal flow in the northwest Gulf of Alaska: The Kenai Current, *J. Geophys. Res.*, *85*(C11), 6680–6688, doi:10.1029/JC085iC11p06680.

Schumacher, J. D., P. J. Stabeno, and A. T. Roach (1989), Volume transport in the Alaska Coastal Current, *Cont. Shelf Res.*, *9*, 1071–1083, doi:10.1016/0278-4343(89)90059-9.

Shcherbina, A. Y., and G. G. Gawarkiewicz (2008), A coastal current in winter: 2. Wind forcing and cooling of a coastal current east of Cape Cod, *J. Geophys. Res.*, *113*, C10014, doi:10.1029/2008JC004750.

Stabeno, P. J., R. K. Reed, and J. D. Schumacher (1995), The Alaska Coastal Current: Continuity of transport and forcing, *J. Geophys. Res.*, *100*(C2), 2477–2485, doi:10.1029/94JC02842.

Stabeno, P. J., N. A. Bond, A. J. Hermann, N. B. Kachel, C. W. Mordy, and J. E. Overland (2004), Meteorology and oceanography of the northern Gulf of Alaska, *Cont. Shelf Res.*, *24*, 859–897, doi:10.1016/j.csr.2004.02.007.

Trenberth, K. E., and J. W. Hurrell (1994), Decadal atmosphere-ocean variations in the Pacific, *Clim. Dyn.*, *9*, 303–319, doi:10.1007/BF00204745.

Weingartner, T. J. (2007), The physical environment of the Gulf of Alaska, in *Long-Term Ecological Change in the Northern Gulf of Alaska*, edited by R. Spies, pp. 12–44, Elsevier, Oxford, U. K.

Weingartner, T. J., et al. (2002), The Northeast Pacific GLOBEC program: Coastal Gulf of Alaska, *Oceanography (Washington, D.C.)*, *15*, 48–63.

Weingartner, T. J., S. L. Danielson, and T. C. Royer (2005), Freshwater variability and predictability in the Alaska Coastal Current, *Deep Sea Res., Part II*, 52, 169–191, doi:10.1016/j.dsr2.2004.09.030.

Williams, W. J., T. J. Weingartner, and A. J. Herrmann (2007), Idealized three-dimensional modeling of seasonal variation in the Alaska Coastal Current, *J. Geophys. Res.*, *112*, C07001, doi:10.1029/2005JC003285.

Wilson, J. G., and J. E. Overland (1986), Meteorology, in *The Gulf of Alaska, Physical Environment and Biological Resources*, edited by D. W. Hood and S. T. Zimmerman, pp. 31–53, Alaska Off., Ocean Assessments Div., NOAA, U. S. Dep. Commer.

Xiong, Q., and T. C. Royer (1984), Coastal temperature and salinity in the northern Gulf of Alaska, 1970–1983, *J. Geophys. Res.*, *89*(C5), 8061–8066, doi:10.1029/JC089iC05p08061.

S. L. Danielson, M. A. Janout, and T. J. Weingartner, Institute of Marine Science, University of Alaska, Fairbanks, Fairbanks, AK 99775, USA. (janout@sfos.uaf.edu)

T. C. Royer, Center for Coastal Physical Oceanography, Old Dominion University, Norfolk, VA 23529, USA.