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Ocean warming and freshening in the northern Gulf of Alaska

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[1] Water column temperatures on the shelf in the northern Gulf of Alaska have increased more than 0.8°C and vertical density stratification has increased since 1970 near Seward, Alaska throughout the 250 m depth. This high latitude marine system has low water temperatures, high rates of precipitation, glacial melting, high wind speeds and high rates of biological productivity. A more than 300 km alongshore shift (locally westward) of isotherms is suggested. The observations are consistent with a conceptual ocean-atmosphere circulation model that employs coastal freshwater discharge, glacial ablation and wind forcing. Positive regional feedback mechanisms accelerate the discharge and poleward heat flux, leading to even higher temperatures, increased ocean stratification and increased storminess. This warming and ocean freshening will have significant impacts on the atmosphere and marine ecosystems of the Northeast Pacific, Bering Sea, Arctic Ocean and quite possibly global ocean circulation. **Citation:** 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.

1. Introduction

[2] High latitudes are reputed to be early and very large responders to climate change. In particular, the very productive marine ecosystem of northern Gulf of Alaska is being subjected to changes that will affect that productivity. Measurements of temperature and salinity acquired in the northern Gulf of Alaska for more than 35 years permit the determination of the response of this system to climate change. Significant temperature changes accompanied by salinity changes are being observed. This is changing the vertical stratification of the water column which alters the flux of nutrients into the euphotic zone from the deep ocean. It also changes the production and retention of phytoplankton and zooplankton in this zone, possibly changing the feeding habits of the fish populations in this region. This is critical to U.S. fisheries production, since in 2004, more than 55% of the U.S. fish landings by weight took place in Alaskan waters (www.st.nmfs.gov).

[3] High latitudes including Alaska are predicted to experience significant climate changes in this century. Climate models predict a warming of 1.5–5°F by 2030 and 5–18°F by 2100 and increases in precipitation of 20–25% in northwestern and northern Alaska [*National Assessment Team*, 2000]. Because of these potentially large ecosystem changes at high latitudes, it is possible that conditions in the North Pacific Ocean are capable of revealing climate changes in their early stages. The potential for a regional atmosphere-ocean-glacier feedback system offers interesting consequences for short and long term climate changes.

[4] The northern Gulf of Alaska contains the largest freshwater discharge system in North America, estimated to be 23,800 $m^3\ s^{-1}$ [*Royer*, 1982] about 62% greater than the Mississippi River discharge (14,700 $m^3\ s^{-1}$) [*Atkinson and Grosch*, 1999]. High rates of coastal precipitation (in excess of 3 m per year) over a relatively narrow (<100 km) drainage area create these high coastal discharge rates. There are no major dams to restrain the runoff and the sparsely populated region limits the accessibility and visibility of this free running freshwater source. The influence of this freshwater on water density in the North Pacific is enhanced through the nonlinear nature of the salinity-temperature-density relationship of seawater [*Millero and Poisson*, 1981] and the relatively low water temperatures. The cross-shelf pressure gradient created by this nearshore discharge creates a buoyancy driven, cyclonic, alongshore geostrophic flow called the Alaska Coastal Current, (ACC) [*Royer*, 1981; *Stabeno et al.*, 2004]. In fall and winter, the atmospheric circulation over the Gulf of Alaska is dominated by the Aleutian Low with easterlies over the northern gulf that cause a coastal convergence with downwelling that constrains the coastal freshwater discharge in the nearshore region [*Reed and Schumacher*, 1986]. This creates speeds in the ACC that can exceed 1.8 $m\ s^{-1}$ [*Johnson et al.*, 1988]. The freshwater discharge is also augmented by glacial meltwater since this coastal region also contains the third largest ice fields in the world [*Meier*, 1984]. Changes in the glaciers have important consequences on the ocean circulation and atmosphere. With recent increases in the glacial melting in Southeast Alaska [*Arendt et al.*, 2002] those glaciers now contribute about 100 $km^3\ year^{-1}$ to the coastal freshwater discharge in the Northeast Pacific and this will enhance the ocean circulation.

2. The Data and Analysis

[5] Long term hydrographic observations in high latitudes are rare but fortunately such measurements have been made at the mouth of Resurrection Bay, Seward, Alaska (60°N, 150°W) in the coastal waters of the northern Gulf of Alaska (Figure 1) [*Royer*, 2005]. These are the northernmost such measurements along the west coast of North America. Water temperature and salinity versus depth (December 1970–present) within the ACC have been assigned to two layers (0–100 m and 100–250 m) in consideration of the mixed layer dynamics (Figure 1). The upper layer captures the major aspects of the mixed layer [*Sarkar et al.*, 2005] and responds seasonally to wind and freshwater discharge [*Royer*, 2005]. The lower layer responds on longer time

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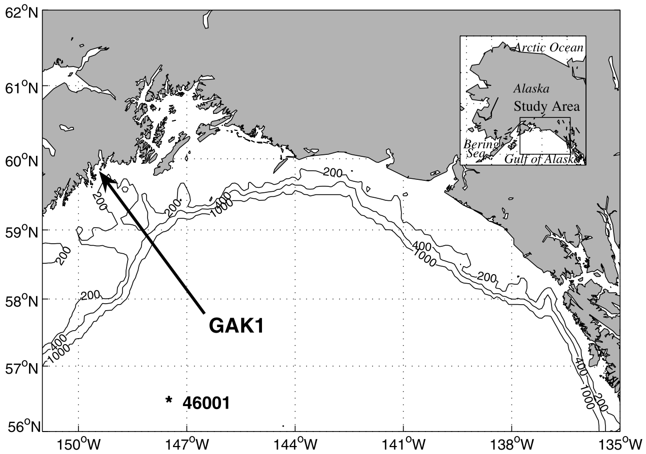

Figure 1. Gulf of Alaska with Station 1 (GAK1) and the location of the National Data Buoy Center (NDBC) Buoy 46001.

scales [*Royer*, 2005]. The measurements have been gathered from many research vessels as they leave and enter the marine research facilities at Seward, Alaska. The temporal sampling intervals are irregular, varying from hours to months. Since September 1990, the sampling has been conducted approximately monthly. The accuracies for temperature, salinity and depth are respectively (±0.01°C), (±0.02 psu) and (±2 m).

[6] Temporal trends of the temperature anomalies were determined for a two layer system (0–100 m and 100–250 m). The monthly means for the entire hydrographic record were determined and removed to yield the anomalies that are departures from these monthly means. An annual sinusoid was determined also but it did not differ significantly from the monthly mean technique.

[7] We first tested for the presence or absence of a trend using Daniells' test, based on Spearman's Rank Correlation [*Kendall and Gibbons*, 1990]. It is a non-parametric technique in which the data are ranked from the algebraically

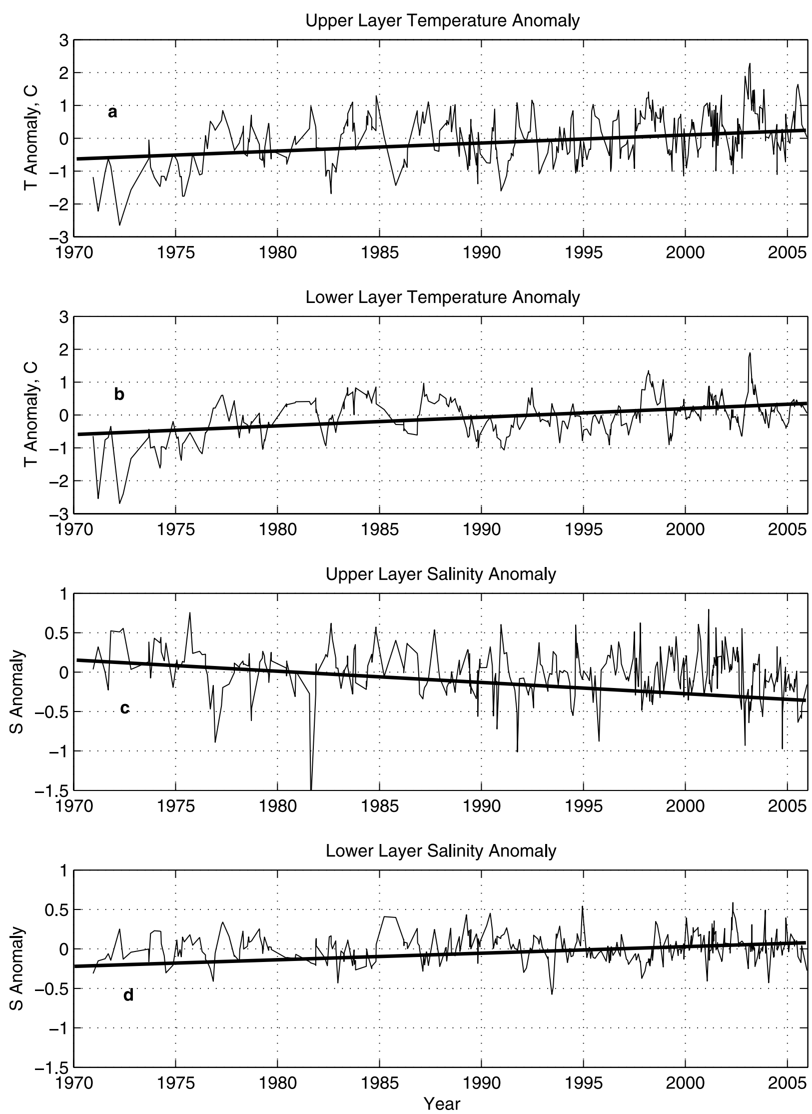

Figure 2. GAK1 anomalies of temperature and salinity for (a) the upper layer (0–100 m), (b) the lower layer (100–250 m), (c) the upper layer (0–100 m) and (d) the lower layer (100–250 m). The respective linear slopes for temperature are +0.031°C/year and +0.024°C/year and for salinity are −0.0040/year and +0.0025/year.

Table 1. Results of Daniells' Test Based on Spearman's Rank Correlation

	1970–2005			1979–2005		
Parameter	N	C	P	N	C	*P*
Upper Layer T	422	+0.329	$<10^{-6}$	373	+0.227	$<10^{-5}$
Lower Layer T	422	+0.308	$<10^{-6}$	373	+0.139	$<10^{-3}$
Upper layer S	422	−0.153	0.001	373	−0.098	0.059
Lower Layer S	422	+0.141	0.004	373	+0.154	0.003

Table 3. Low Pressure Events for NOAA Buoy 46001 (56.30°N, 148.17°W)

Period	Number of Months	Mean Number of "Low Pressure Events" Per Month
11/74–07/78	38	4.37
03/79–12/90	131	4.65
06/94–06/96	25	4.20
11/97–12/05	98	5.89

smallest to largest values. The values of the ranks are correlated with the corresponding times. The result of the test is a pair of values, the correlation coefficient C and the probability, P, that the correlation is due to chance. Positive (negative) values of C indicate an increasing (decreasing) trend and the magnitude of C is a rough indicator of the relative magnitude of the slope.

[8] We applied this test to the time series of the temperature and salinity data plotted in Figure 2. Also, noting that the early 1970s were quite cold, and that this might bias the results we applied the test to the same data for the period 1979–2005. The results of these calculations, with N the number of data points, are presented in Table 1.

[9] For both time periods, the positive values of C for the temperature time series indicate an increasing trend with extremely small probability that this is due to chance. The upper layer salinity time series has a negative correlation showing that the salinity is decreasing. However, while the upper layer salinity P for the 1970–2005 is quite small, for 1979–2005 it is barely significant thus suggesting that the decrease in salinity in the upper layer is rather small. In contrast, the salinity in the lower layer has a positive value of C for both periods and the corresponding low values of P indicate that this is significant.

[10] To quantify the slopes of the trends we did a least squares fit to the anomalies and found the 95% and 90% confidence intervals. These confidence intervals were calculated in the standard way first ignoring the autocorrelation of the time series, i.e., assuming that the integral scale I = 1.

[11] In order to obtain an estimate of the integral scale we must have the data evenly spaced in time so that the autocorrelation functions can be calculated. These data were interpolated onto monthly time series from January 1971 to December 2004 giving 408 values. Then the effective N = N/I. A least squares fit to the interpolated temperature and salinity anomalies was done and the 95% and 90% confidence intervals were computed in the standard manner using the effective N = (N/I) (Table 2).

Table 2. Linear Fits of Original Data and Interpolated Data Using Effective Sample Size

Original Data	I	N	Slope	95% C.I.	90% C.I.
Upper Layer T	1	422	+0.031	±0.007	±0.006
Lower Layer T	1	422	+0.024	±0.006	±0.005
Upper Layer S	1	422	−0.0040	±0.0029	±0.0024
Lower Layer S	1	422	+0.0025	±0.0017	±0.0014
Interpolated Data	I	N/I	Slope	95% C.I.	90% C.I.
Upper Layer T	24	17	+0.039	±0.035	±0.029
Lower Layer T	28	14	+0.031	±0.034	±0.028
Upper Layer S	4	102	−0.0052	±0.0052	±0.0044
Lower Layer S	2	204	+0.0015	±0.0024	±0.0020

[12] The slopes of the fits to the temperature anomalies are significant with C.I. > 95% for the original data and with C.I. > 90% for the interpolated data. There is an increase in temperature of +0.031°C year^{-1} in the upper layer and +0.024°C year^{-1} in the lower layer (Figures 2a and 2b) with slightly larger slopes for the interpolated data. Thus, over the 35 year record, water temperature anomalies have increased by about 1°C. These temperature anomalies do not propagate downward from the sea surface [*Royer*, 2005]. For this same period, the salinity anomalies decreased in the upper layer at a rate of −0.0040 year^{-1} (Figure 2c) and increased in the lower layer at a rate of +0.0025 year^{-1} (Figure 2d). The relatively noisy signals (Figure 2) are expected since the extraction of the relatively large seasonal signal has been predicted to lead to decadal and interdecadal variability in climate signals [*Huybers and Curry*, 2006]. The contrasting trends in the upper and lower layer salinities increase the overall vertical density stratification during this period. Increased stratification will reduce the vertical circulation and will inhibit the penetration of incoming surface heat fluxes. This is consistent with the lack of vertical propagation of the temperature anomalies.

[13] The upper layer temperature change at GAK1 (+0.031°year^{-1}) is 1.5 to 5 times the rates reported earlier by *Freeland et al.* [1997] for the sea surface at Ocean Station P and coastal British Columbia. The GAK1 upper layer salinity change (−0.004 year $^{-1}$) is about the same as *Freeland et al.* [1997] found for the sea surface at Ocean Station P but half to ¼ of the rate for coastal British Columbia. They used the time period from 1956 to 1995 for the Ocean Station P observations and from about 1935 to 1995 for the B.C. coastal measurements. They also concluded that the density of the upper layer was decreasing and vertical stratification was increasing.

[14] Increased stratification will enhance the horizontal pressure gradients and horizontal circulation because the influx of freshwater (decreased salinity) is not uniform across the shelf. The freshwater discharge at the coast enhances the horizontal pressure gradient across the shelf and accelerates an alongshore, westward (large scale, cyclonic) current here. The ocean temperature gradient is positive toward the east in the Gulf of Alaska and ultimately equatorward (www.nodc.noaa.gov/OC5/WOA01/pr_woa01.html) so that the alongshore acceleration will bring water with higher temperatures into the region. A typical alongshore thermal gradient below the mixed layer is 0.0026°C km^{-1} [*Royer*, 2005]. Note: this horizontal temperature gradient is an order of magnitude less than that originally reported by *Royer* [2005]. A diligent reviewer of this manuscript noted the error and our recalculation of the

Table 4. Low Pressure Event Probabilities

Period	Probability of Equality of Means		
	03/79–12/90	06/94–06/96	11/97–12/05
11/74–07/78	0.471	0.467	0.005
03/79–12/90		0.244	0.021
06/94–06/96			0.003

gradient revealed the mistake in that earlier paper. The thermal contours in the lower layer, that are generally perpendicular to the coast, could shift 314 km alongshore (northward) since 1970 as a result of the increased transport in the ACC. Assuming that these alongshore temperature gradient in the Gulf of Alaska is constant (0.0026°C km^{-1}) over the period of the record and that the observed temperature changes are only caused by advection, then an increase in the current of about 3×10^{-4} m s^{-1} would be required to move a given isotherm 314 km alongshore over 34 years. If a barotropic, geostrophic coastal current is assumed, the cross shelf slope change would be 4×10^{-9}. A change in the sea level at the coast would be about 8×10^{-4} m for a current width of 200 km or 8×10^{-5} m for a width of 20 km. Either of these changes would be difficult to detect in coastal sea level measurements since the isostatic rebound of the sea level changes along this coast is of the order of 0.15 m per 30 years (5×10^{-3} m $year^{-1}$ [*Lisitzin*, 1974, p. 175]).

[15] In addition to the enhanced alongshore flow; there is an enhanced cross-shelf estuarine-type flow as suggested by *Tully and Barber* [1960]. Since the cross-shelf thermal gradients are much less than the cross-shelf salinity gradients [*Reed and Schumacher*, 1986], increases in the lower layer salinity anomaly are expected with enhanced estuarine-type flow.

[16] The observed changes in this two layer system are consistent with a conceptual atmosphere-ocean feedback model [*Royer et al.*, 2001]. The increase in the freshwater discharge decreases the upper layer salinity, accelerating the alongshore flow. The accompanying enhanced alongshore (poleward) heat transport increases the ocean temperature. This increases cyclogenesis (storm activity), reinforcing the poleward (cyclonic) transport. Increased cyclogenesis should accompany the increased coastal discharges and heat transport. This would add to the precipitation and alongshore wind driven transports in this positive feedback air-sea system.

[17] Increased storminess is seen in the sea level barometric pressures (slbp) since 1974 at 56.30°N 148.17°W (NOAA Buoy 46001) (Figure 1). Hourly values of slbp are available from November 1974 to December 2005 with a number of substantial time gaps. During time periods without such gaps, there are sometimes missing hourly data. For example, in 2005, there were 10 hourly values missing out of a possible 8760 hours. In these cases, the missing values were replaced by linearly interpolated values. Using these more complete data, we computed daily averages of the slbp. We defined a day as having a "low pressure event" (a storm) if the mean slbp for that day was less than or equal to 995 hPa. For four periods since 1974, as determined by major data gaps, we counted the number of "low pressure events" and computed the mean number per month (Table 3).

[18] The data suggest that there has been an increase in the mean number of "low pressure events" per month in the most recent period (11/97–12/05) as compared to each of the three other periods for which data are available. To test the conclusion that there has been such an increase, we used a t-test to calculate the probability that the means are equal for each pair of periods (Table 4).

[19] The number of "low pressure events" per month do not differ significantly between the three earliest periods. However, the most recent period has had significantly more "low pressure events" than any of the preceding periods.

[20] These increases in the cyclogenesis are consistent with the conceptual feedback model for the Gulf of Alaska [*Royer et al.*, 2001]. An enhanced coastal precipitation along with the increased storminess work together to positively reinforce increased poleward heat transport by the ocean.

[21] Globally, increased heat transports will cause climate changes that influence glacial volumes in Greenland [*Rignot and Kanagaratnam*, 2006] and Alaska [*Arendt et al.*, 2002] that in turn affect the ocean volumes and salinities. The Alaska mountain glaciers in this coastal drainage are significant contributors to the freshwater discharge and changes in their volumes must be considered. Increases in coastal freshwater discharge since 1970 [*Royer*, 2005] are consistent with the increases in the water temperatures, decreases in the upper layer salinity and accelerations within the ACC. The original discharge model of *Royer* [1982] assumes that over the duration of that particular modeling effort (1931–1980) there were no net changes in the volume of the glaciers. Thus, all of the precipitation eventually reached the ocean, though interannual changes in glacial volume were allowed depending of the regional air temperatures and precipitation. However, recent glacial studies by *Arendt et al.* [2002] have provided estimates of the changes in glacial volumes. From 1980 to 1995, the glacial volume in the Alaska coastal mountain drainages shrank by 52 km^3 $year^{-1}$ [*Arendt et al.*, 2002]. Since 1995 the glacial ablation has nearly doubled to 96 km^3 $year^{-1}$. These increases in the discharge rate due to glacial ablation are 271 m^3 s^{-1} $year^{-1}$ (1980–1995) and 438 m^3 s^{-1} $year^{-1}$ (since 1995). (The percentage of these volume changes are similar to decadal changes in the Greenland Ice Sheet that increased from 90 to 200 km^3 $year^{-1}$ [*Rignot and Kanagaratnam*, 2006]).

[22] The rate of increase of the temperature at GAK1 of about 0.03°C $year^{-1}$ throughout the water column is very similar the rate of increase in the sea surface temperature of 0.04°C $year^{-1}$ at Woods Hole, MA for the period 1970–2002 [*Nixon et al.*, 2004]. However, the behavior of the salinity-temperature anomalies in the North Atlantic is in sharp contrast to those found in this study for the North Pacific. In the North Atlantic features such as the Great Salinity Anomaly [*Dickson et al.*, 1988] have low salinities associated with low temperatures. Such a relationship is a result of the surface waters having their origin in the Arctic with its relatively low temperatures. Contrary to the North Atlantic Ocean, in the Northeast Pacific Ocean, the lowered salinities are associated with increased temperatures.

[23] The rate of the temperature increase in the Gulf of Alaska (0.3°C per decade) is considerably higher than the rate of 0.06°C per decade (0.31°C from 1948 to 1998) for the global ocean for the upper 300 m reported by *Levitus et*

al. [2000]. The higher rate for the Gulf of Alaska could be caused by the regional influences of the ocean-glacial-atmosphere interactions.

[24] The changes in temperature and salinity at this hydrographic station could be indicators of wider, regional changes. Concurrent changes in the hydrography in the Gulf of Alaska, Bering Sea and Bering Straits suggest that teleconnections play an important role. For example, increases in the upper layer water temperatures and decreases in the upper layer salinities over the past several decades in the Gulf of Alaska near Seward, coincide with temperature increases at the mooring M2 in the Bering Sea [*Overland and Stabeno*, 2004] and temperature increases and salinity decreases at a mooring in the Bering Straits [*Woodgate and Aagaard*, 2005].

[25] How will these changes in the regional hydrography, in general, and local hydrography, more specifically, affect the primary production in on the eastern Bering Sea shelf? The water column stratification will change, probably increasing the stability since both warming and freshening will enhance the stability. However, the possible absence of sea ice will diminish the vertical stratification since the extrusion of salt from the ice during the freezing process will not occur and the springtime melting will not occur. Both of these processes would add stability to the spring water column. These changes and the timing of the seasonal cycles might change the spring bloom. However, a major factor in the character of the water column will be the wind mixing and must be considered, as well.

3. Conclusions

[26] The processes affecting the upper layer temperatures and salinities in the Northeast Pacific could have even farther impacts. *Seidov and Haupt* [2005] emphasize the interconnections between the North Pacific and North Atlantic Oceans. They propose that the driving of the global ocean "salinity conveyor belt" is related not just to the salinities of the upper waters of the North Atlantic but also to the salinities of the upper waters of the North Pacific, more specifically on the differences between the salinities of these two regions. Predictions of diminished deep water convection in the North Atlantic resulting from decreased upper layer salinities might be incorrect since freshening in the North Pacific might be enhancing the overall thermohaline circulation [*Seidov and Haupt*, 2005]. It is important to consider the freshwater disparity between the North Atlantic and North Pacific in determining the changes in the global ocean circulation.

[27] Increases in temperature, freshwater discharges, storminess, lower layer salinity and decreased upper layer salinity are consistent with a conceptual feedback model [*Royer et al.*, 2001]. The implications of this model are that regional ocean-glacial-atmosphere processes will reinforce one another leading to further, possibly accelerated changes in the future. Increased stratification of the upper layers in the Gulf of Alaska will continue.

[28] Increased advection will shift the coastal ecosystem more than 300 km northward and westward around the Gulf of Alaska. This shift in combination with extremely sharp thermal limits of some marine mammal and fishes such as sockeye salmon in the North Pacific [*Welch et al.*, 1998] might shift those populations out of the Gulf of Alaska into the Bering Sea. Stratification increases will have impacts on entire marine ecosystem. It controls the flux of nutrients into the euphotic zone and the strength (depth gradient of density) and depth of the mixed layer [*Sarkar et al.*, 2005]. Nevertheless it is uncertain as to whether the changes reported here will increase or decrease the mixed layer depth since the vertical density contrast would tend to decrease this depth while increased storminess would increase the depth through increased wind mixing. *Polovina et al.* [1995] observed an increase in the mixed layer depth in the Central and North Pacific during the 1977–88 period in comparison with 1960–76 and attributed the changes to increased storminess. However, they used only temperature profiles for the mixed layer determination so their conclusions might not be valid for the high latitude North Pacific. As a final point, changes in salinity of the Northeast Pacific will affect the salinity in the Bering Sea and Arctic Ocean and could impact the global "salinity conveyor belt" [*Seidov and Haupt*, 2005], that would cause this regional circulation change to have global implications.

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