Low-Frequency Pycnocline Variability
in the Northeast Pacific

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Abstract

We use the output from an ocean general circulation model (OGCM) driven by observed surface forcing to examine the upper ocean changes that took place in the Gulf of Alaska over the period 1958-1997, including the 1976-77 climate regime shift. The pycnocline deepened after the mid-seventies in a broad band along the coast, and shoaled in the central part of the Gulf of Alaska. The changes in pycnocline depth diagnosed from the model are in agreement with the pycnocline depth changes observed at two ocean stations in different areas of the Gulf of Alaska. Using a simple Ekman pumping model with linear damping we show that a large fraction of pycnocline variability in the OGCM can be explained by local Ekman pumping. The fit of the simple model to the OGCM is maximized in the central part of the Gulf of Alaska, where the pycnocline variability produced by the simple model can account for ~70-90% of the pycnocline depth variance in the OGCM. Evidence of westward propagating Rossby waves is found in the OGCM, but Rossby waves do not seem to be a dominant signal. On the contrary, large-scale pycnocline depth anomalies have primarily a standing character, thus explaining the success of the local Ekman pumping model. The agreement between simple model and OGCM deteriorates in a large band along the coast, where propagating disturbances within the pycnocline appear to play an important role in pycnocline variability. Coastal propagation of pycnocline depth anomalies is especially relevant in the western part of the Gulf of Alaska, where local Ekman pumping changes would lead to pycnocline variations of the opposite sign than those in the OGCM. The inclusion of a propagation term in the simple Ekman pumping model considerably improves the agreement with the OGCM along the coast. The propagation speed estimated from the OGCM is consistent with mean flow advection in the western part of the coastal path, but is much larger than the OGCM advective velocity in the east, suggesting that wave propagation may also be involved in that area.
1. Introduction

The oceanic pycnocline, the subsurface layer characterized by large vertical density gradients, can be viewed as the interface between the surface ocean mixed layer and the deep ocean. Changes in pycnocline depth may be indicative of changes in upwelling, a process that influences the exchange of properties between the deep and upper ocean. Large areas of the northeast Pacific are characterized by a fresh and well-mixed surface layer which is separated from the deeper ocean by large salinity gradients, or halocline. At high-latitudes, where temperatures are low, salinity has a dominant influence on density, particularly in winter (Freeland et al. 1997), as illustrated by the seasonal evolution of upper-ocean temperature, salinity and density at the location of station Papa (50°N, 154°W) in an ocean general circulation model (OGCM, Figure 1). We can notice in Figure 1 the progressive erosion of the seasonal pycnocline from summer to winter, and the controlling role of salinity upon the vertical stratification in winter. During the winter season, the pycnocline lies at the base of the mixed layer, so that pycnocline depth is very similar to mixed layer depth. Thus, understanding the processes governing pycnocline variability can also help understand the changes in mixed layer depth, a quantity that has a large influence upon biological productivity (Polovina et al. 1995, Gargett 1997).

Pycnocline depth changes result from the local response to surface atmospheric forcing combined with depth anomaly propagation through advection or wave processes. Unlike sea surface temperature (SST) variations, which are largely affected by high-frequency air-sea interactions, pycnocline depth changes are more representative of the low-frequency oceanic response to atmospheric forcing, and can more easily be used to detect climate variations at interannual and decadal timescales. A well-known example of Pacific “decadal” variability is the climate shift that started in late 1976, and was associated with large-scale atmospheric and oceanic changes, including a deepening of the Aleutian Low pressure system, a decrease in SST in the central Pacific and an increase in SST in the eastern Pacific, as well as basinwide changes in a large set of environmental and biological quantities (Ebbesmeyer et al. 1991).
Using subsurface observations over the period 1968-1990, Lagerloef (1995) examined interannual and decadal variations of the dynamic height field, which is closely related to pycnocline topography, over the Gulf of Alaska. The dominant mode of variability of the dynamic height field, identified through empirical orthogonal function (EOF) analysis, captured the 1976-77 climate shift, and was well correlated with other climate indices. A simple model for the evolution of the dynamic height field, in which the forcing was supplied by the local Ekman pumping, and dissipation was modeled with a linear damping term, appeared to reproduce a large fraction of the low-frequency dynamic topography variations. However, the hindcast of dynamic heights underestimates the observed variability in the western half of the gyre, a result Lagerloef (1995) attributes to the presence of westward propagating baroclinic Rossby waves.

Cummins and Lagerloef (2002, CL02 hereafter) used the same Ekman pumping model described by Lagerloef (1995) to examine the structure of pycnocline depth variability that is excited by the dominant patterns of anomalous Ekman pumping over the northeast Pacific (30°-60°N, 180°-120°W). The evolution of pycnocline depth predicted by the simple Ekman pumping model reproduced well the observed variations at Ocean Weather Station Papa (OWS P, 50°N, 145°W). However, CL02 were unable to test the performance of the simple model at other locations in the Gulf of Alaska, and to verify the accuracy of the spatial patterns of pycnocline depth changes produced by the model over the entire northeast Pacific.

The success of the simple Ekman pumping model in explaining pycnocline variability is somewhat surprising given our understanding of the ocean adjustment to changes in surface forcing through Rossby wave propagation. Satellite altimetry shows, indeed, westward propagation of sea surface height (SSH) anomalies in the northeast Pacific (Kelly et al. 1993, Fu and Qiu 2002), and Qiu (2002) showed that the SSH variability from the Topex/Poseidon (T/P) altimeter mission (1993-2001), areally averaged over the offshore region of the Gulf of Alaska, could be largely explained by first-mode baroclinic Rossby wave propagation. Comparison of the performances of a 1 1/2 quasigeostrophic model (which includes Rossby wave propagation) and the local Ekman pumping model of CL02 in explaining SSH variability in the northeast Pacific (Cummins and
Lagerloef 2004, CL04 hereafter) indicates that the hindcast skill of the simple model is very similar to that of the quasigeostrophic model. Thus, CL04 concluded that interannual variability in the Gulf of Alaska is dominated by the local response to wind forcing.

In this study we extend the work of CL02 and CL04 by considering the relative role of Ekman pumping and oceanic processes in producing low-frequency (interannual and decadal) pycnocline variations over the northeast Pacific during 1958-97. The specific questions we ask are: what fraction of pycnocline variability can be explained by local Ekman pumping in different areas of the Gulf of Alaska? If the simple model of CL02 fails in some areas, what other processes control pycnocline changes in those areas? To answer these questions we will use the output from an ocean general circulation model (OGCM) driven by observed surface forcing. After testing the model’s performance at some locations where long-term observations are available, the OGCM output is compared to the simple model of CL02, and then used to examine the influence of oceanic processes, i.e. advection and wave propagation, upon pycnocline variability in different areas.

In section 2 we describe the OGCM, in section 3 we compare the pycnocline depth changes in the OGCM with those observed at two oceanographic stations in different areas of the Gulf of Alaska, in section 4 we compare the pycnocline variability predicted by the simple model with that obtained from the OGCM, and in section 5 we examine the influence of Rossby wave propagation on low-frequency pycnocline variations. We will find that the simple model can account for a large fraction of the OGCM pycnocline variability in the central part of the Gulf of Alaska, but its skill decreases in a broad band along the coast, especially in the area of the Alaskan Stream. Advection and boundary wave propagation may be very important processes in the coastal regions. In section 6 we examine the role of these processes by extending the simple model to include propagation of pycnocline depth anomalies along the coast. In section 7 we examine the validity of the Sverdrup balance at decadal timescales. We conclude in section 8.
2. The OGCM

The OGCM used for this study is the NCAR ocean model (NCOM) that has been described in detail by Large et al. (1997), Gent et al. (1998), and Large et al. (2001). The specific numerical simulation analyzed here is described in Doney et al. (2003). In this section we only provide a brief summary of the basic model characteristics and information about the surface forcing used for this simulation.

NCOM is derived from the Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model with the addition of a mesoscale eddy flux parameterization along isopycnal surfaces (Gent and McWilliams 1990) and a nonlocal planetary boundary layer parameterization (Large et al. 1994). The model is global, with a horizontal resolution of 2.4° in longitude and varying resolution in latitude, ranging from 0.6° near the equator to 1.2° at high-latitudes. The model version used for this study includes an anisotropic viscosity parameterization (Large et al. 2001) with enhanced viscosity close to ocean boundaries and much weaker viscosity in the ocean interior.

The surface forcing includes momentum, heat and freshwater fluxes for the period 1958-1997. The wind stress is computed from the reanalyses fields produced at the National Center for Environmental Prediction (NCEP; Kalney et al. 1996) using bulk formulae. The sensible and latent heat fluxes are computed from the NCEP winds and relative humidity and the model’s SSTs using standard air-sea transfer equations (Large and Pond 1982, Large et al. 1997). Sensible and latent heat fluxes depend on the difference between SST and surface air temperature. Since SST and air temperature closely track each other, when observed air temperatures are used in the bulk formulas, as in the present model simulation, the model’s SST is relaxed toward observations (Haney 1971). The relaxation timescale is relatively short (30-60 days for typical mixed layer depths), so that the SST in the model can be expected to be strongly constrained by the surface forcing rather than by the interior ocean dynamics.

The numerical simulation is started from an initial condition obtained from a preliminary climatological integration, so that the initial model state is not too different from the mean state characteristic of the 40-year experiment. Then the model was run for two 40-year cycles, the sec-
ond cycle starting from the conditions achieved at the end of the first 40-year segment. The mismatch between the model state and the forcing at the beginning of the second cycle did not seem to produce any long-term transient behavior, but some residual drift in temperature and salinity can be detected at depths larger than approximately 500 m (Doney et al. 2003). Here we analyze the output for the second 40-year period using monthly mean values.

3. How realistic is the pycnocline variability in the OGCM?

Lagerloef (1995) has shown that a significant fraction (28%) of the subsurface variance over the period 1960-1990 is associated with the 1976-77 climate regime shift. Thus, we start by examining the pycnocline depth variations exhibited by the OGCM in association with the 1976-77 climate shift. Changes in the density structure are detectable to depths greater than ~800 m, but appear to be more pronounced in the upper ocean, as shown by a zonal density section along 53.8°N (Figure 2). Above 400m, the isopycnals deepen toward the eastern and western edges of the gyre after the mid-seventies, resulting in a more pronounced doming of the density structure. The depth of the $26.4\sigma_\theta$ isopycnal, which lies in the core of the main pycnocline, is used as proxy for pycnocline depth. The mean depth of the $26.4\sigma_\theta$ isopycnal ranges from ~100 m in the center of the Alaska gyre to ~150-180 m around the rim of the gyre. The changes in pycnocline depth are computed as the difference between the pycnocline depth in the period 1977-1997 (Period 2) and the pycnocline depth in the period 1960-1975 (Period 1). After the mid-seventies, the pycnocline was shallower in the central part of the Gulf of Alaska, and deeper in a broad band following the coast (Figure 3). The deepening was more pronounced in the western part of the Gulf of Alaska, to the southwest of Kodiak Island (“K” in Figure 3), following approximately the Alaska Peninsula. A similar pattern of pycnocline changes was found by Miller et al. (1994) in a coarser resolution (~4°) ocean model hindcast, using surface forcing fields from the COADS dataset instead of NCEP. The leading mode of a simple stochastic model forced by Ekman pumping also displayed the spatial structure and temporal evolution characteristics of the 1976/77 regime shift (Figure 4 in CI02). Is the pattern of pycnocline variations in Figure 3 realistic?
Subsurface observations are generally sparse, and salinity data in particular are very limited. Since salinity is an important factor controlling the potential density distribution in the subarctic Pacific, the paucity of salinity data limits the possibility of testing the performance of the OGCM over the whole northeast Pacific. However, there are a few oceanographic stations in the Gulf of Alaska, where measurements of both temperature and salinity are available over a long period of time including the 1976-77 climate shift. We will use measurements from two stations, Ocean Weather Station Papa (OWS P, “P” in Figure 3), which is located in the area where the pycnocline becomes shallower, and station GAK1 (“G” in Figure 3) which is in the band where the pycnocline deepens.

OWS P (50°N, 145°W) provides a long records of bottle cast and CTD data over the period 1957-94. The time series has been recently augmented by more recent observations from 1995 to 1999. Generally, the density of observations is higher in the earlier period, 1957-1981, when station P was occupied on a regular basis by a weather ship. Freeland et al. (1997) have computed pycnocline depths by assuming a 2-layer representation of the density structure and using a least squares approach to determine the thickness of the upper layer from the observed density profiles. The 2-layer representation is appropriate for wintertime (December - April) conditions, and it tracks the core of the main pycnocline in winter. CL02 have used the additional observations after 1994 to extend the time series of mean winter pycnocline depth from 1994 to 1999. Thus, the whole time series covers the period 1957-1999, with gaps in 1985, 1988, 1990, and 1994. The resulting time series is compared in Figure 4a with the winter pycnocline depth from the OGCM. The correlation between the two time series is 0.7, and the amplitude of the response is also approximately correct. Notice that the observed pycnocline depth anomaly values are often averages of few data over the winter months, while the winter pycnocline depths estimated from the OGCM are based on monthly averages of potential density which is available at each model time step (of the order of one day). Thus, higher-frequency processes (e.g. mesoscale eddies and baroclinic tides) which are filtered out in the OGCM, may contribute to some of the observed pycno-
cline variability. Both model and observations indicate a shoaling trend of the main pycnocline, as discussed by Freeland et al. (1997), and CL02.

Station GAK1 (59°N, 149°W) is located at the mouth of Resurrection Bay near Seward, Alaska, where the OGCM indicates a deepening of the pycnocline after the mid-seventies (Figure 2). Profiles of temperature and salinity to a depth of 250 m have been measured starting in December 1970, and are still ongoing. The number of observations over the winter period varies from less than five measurements in 1972, 1980, 1981, and 1985 to a maximum of 15 values in 1999. Starting from the individual density profiles, we have estimated the depth of the 25.4 $\sigma_\theta$ isopycnal, which is located close to the base of the observed winter pycnocline. Winter values have then been computed by averaging all the values between December and April. Standard deviations about each winter value (not shown) can be as large as 50 m. Station GAK1 is situated very close to shore, and there is no model grid point available at the exact location of the station. So we have estimated winter pycnocline depths from the OGCM at the closest model grid point (58.6°N, 147.6°W). The comparison between the observed and simulated depth anomalies is shown in Figure 4b. Although the observed time series exhibits interannual fluctuations of larger amplitude than the time series from the OGCM, which are based on monthly averages of values at each model time step, the two time series agree quite well (the correlation coefficient is 0.77) on both interannual and longer timescales (apart from a discrepancy in 1979-81). In particular, the low-frequency behavior is very similar in both time series, and indicates a deepening of the pycnocline after the mid-seventies. Thus, the comparison with observations indicates that the pattern of pycnocline depth changes produced by the OGCM is believable. We then use the OGCM to assess which fraction of the pycnocline variability can be explained by local Ekman pumping.

4. Where is Ekman pumping important?

The Ekman pumping is defined as the vertical component of the curl of the wind stress $\mathbf{\tau}$, divided by the Coriolis parameter $f$ and the mean density of sea water $\rho_o$:

\[
\mathbf{\tau} = \frac{1}{\rho_o} \left( \left( \frac{\partial u}{\partial y} \right) \mathbf{j} - \left( \frac{\partial v}{\partial x} \right) \mathbf{i} \right)
\]
The Ekman pumping field is influenced by high-frequency weather fluctuations, and contains a large fraction of energy at periods shorter than a few months. It also contains slow climate fluctuations. The changes in the Ekman pumping field associated with the 1976-77 climate shift are shown in Figure 5, as the difference between Ekman pumping during winter (December-April) in Period2 (1977-97) and Period1 (1960-75). The resulting pattern shows negative $W_E$ changes in a broad band that follows the coast to the east of Kodiak Island (indicated with a “K” in Figure 3), and positive changes elsewhere in the Gulf of Alaska, with maximum values extending to the southeast of the Alaskan Peninsula. The largest negative difference is achieved in the very northern part of the Gulf of Alaska, between approximately Seward (“S” in Figure 3) and Queen Charlotte Island (“Q” in Figure 3). Since the time averaged Ekman pumping is positive over the whole Gulf of Alaska, a positive (negative) difference implies that $W_E$ is more (less) favorable to upwelling after the 1976-77 climate shift. The pattern of $W_E$ changes in Figure 5 is very similar to that computed by Lagerloef (1995) using an empirical orthogonal function (eof) approach. Lagerloef (1995) found that ~87% of the variance of the $W_E$ field over the area 50°-60°N, 160°W-130°W is explained by the first two eofs. In particular, EOF 2 reflects the slow variations associated with the 1976-77 climate shift, and has a spatial structure very similar to the pattern in Figure 5.

The pattern of $W_E$ changes is consistent with the distribution of pycnocline changes in Figure 3: areas of negative $W_E$ differences generally correspond to areas of deeper pycnocline during Period2, and vice versa. The major discrepancy between the two fields occurs in the band extending southwestward from Kodiak Island to the Aleutian Islands, where a positive change in $W_E$ from Period1 to Period2 is associated with a deeper pycnocline. The local correlation between Ekman pumping and pycnocline depth in the OGCM ($h_{OGCM}$) is computed after low-pass filtering the Ekman pumping to remove periods shorter than 2 years, and considering the maximum correlation as a function of lag between $W_E$ and $h_{OGCM}$. The correlation is maximized in a band
that extends westward from the center of the Alaskan Gyre (~50°N, 145°W) to the Aleutian Islands, with values of ~0.7 (Figure 6a). The correlations decrease toward the Alaska Peninsula, where they drop to values very close to zero. Along the coastal band to the east of Kodiak Island, correlations tend to be higher (0.3-0.5) than on the western side, with maximum values of ~0.5 between Seward and Queen Charlotte Island. Figure 6b shows the lags between Ekman pumping and pycnocline depth that yield the maximum correlations (Figure 6a). Ekman pumping variations always lead pycnocline depth variations with lags ranging from 6-8 months over a large area in the central part of the Gulf of Alaska to values as large as 10-15 months in the northern portion of the Gulf of Alaska, and close to the southeastern corner of the domain. Lags are much smaller from the southern part of the Alaska Peninsula to the Aleutian Islands, where Ekman pumping and pycnocline depth are very little correlated.

To further examine the influence of the local Ekman pumping upon pycnocline variability and understand the phase relationships between the two fields, we consider the simple model used by Lagerloef (1995) and CL02:

$$\frac{dh}{dt} = -W_E - \lambda h,$$  \hspace{1cm} (2)

where $h$ is the pycnocline depth, $W_E$ the Ekman pumping, and $\lambda$ a linear damping coefficient with the units of the inverse of time. Equation (2) is forced with monthly anomalies of Ekman pumping computed from the NCEP/NCAR reanalyses wind stress, the same forcing which drives the OGCM. Equation (2) is solved using a second-order accurate trapezoidal scheme as

$$h^{n+1} = \alpha_1 h^n + \alpha_2 W_E^{n+\frac{1}{2}},$$  \hspace{1cm} (3)

where $\alpha_1 = (2 - \lambda \Delta t)/(2 + \lambda \Delta t)$ and $\alpha_2 = (2 \Delta t)/(2 + \lambda \Delta t)$. We have used the OGCM pycnocline depth anomalies at the initial time (January 1958) as initial condition for the simple model.
A question remains concerning the value of the damping timescale $\lambda^{-1}$. CL02 found that the time evolution of the pycnocline depth produced by the simple model ($h$) is largely controlled by the damping timescale $\lambda^{-1}$. Longer damping timescales tend to emphasize the oceanic response at lower frequencies, and to increase the lag between Ekman pumping and pycnocline depth variations. Using a linear fit to the dynamic height field, Lagerloef (1995) found that appropriate damping timescales are in the range 0.5-2.5 yr, with a median value of about 1 yr. As shown by (3), $\alpha_1$ can be interpreted as the lag 1 autocorrelation of pycnocline depth, so that $\lambda^{-1}$ could be estimated from observed values of lag 1 autocorrelations. Using the lag 1 autocorrelation of yearly pycnocline depth anomalies at OWS P CL02 found a value of 1.4 yr for $\lambda^{-1}$, while the lag 1 autocorrelation of the Pacific Decadal Oscillation (PDO) index (the principal component of the leading EOF of monthly SST anomalies over the North Pacific, as computed by Mantua et al. 1997) yielded $\lambda^{-1} = 1.6$ yr.

CL02 used a reference value of 1.5 years, which was assumed to be constant over the domain. However, if the damping term $-\lambda h$ in (2) is a parameterization of oceanic processes, the damping timescale can be expected to vary spatially, depending on the strength of those processes. Using the OGCM as a term of comparison, we have solved equation (2) at each model grid point for values of $\lambda^{-1}$ ranging from 0.5 years to 10 years at intervals of 0.5 years, and have determined the value of $\lambda^{-1}$ that maximizes the instantaneous correlation with the OGCM time series at each grid point. The resulting spatial distribution of $\lambda^{-1}$ (Figure 7a) shows indeed a large spatial variability of the damping timescale, ranging from 14-16 months in the central part of the Gulf of Alaska, to values as large as 90 months along the coastal band to the east of Kodiak Island. In the area along the Alaska Peninsula, from Kodiak Island to the Aleutian Islands, $\lambda^{-1}$ is close to 12 months, but correlations are very low in this area (Figure 7b). The values we find for the damping timescale in the central part of the Gulf of Alaska (14-18 months) are in agreement with Lagerloef (1995) and CL02’s estimates.

The pattern of $\lambda^{-1}$ in Figure 7a is generally consistent with the spatial distribution of the lags between $W_E$ and pycnocline depth in Figure 6a. From equation (2) the phase lag between $h$ and
$W_E$ is frequency dependent and given by $\Phi = \text{atan}(\omega \lambda^{-1})$, where $\omega$ is the angular frequency. Thus, $\Phi$ depends upon the ratio between the frictional timescale $\lambda^{-1}$ and the period of the forcing $2\pi \omega^{-1}$, varying from values very close to 0°, when the period of the forcing is much longer than the frictional timescale, to $\pi/2$, for forcing periods much shorter than the frictional timescale. Thus, at low-frequencies Ekman pumping and pycnocline depth variations tend to be in phase, while at high-frequencies they are in quadrature. Figure 6a shows that the lag between Ekman pumping and pycnocline depth response is generally larger in areas where $\lambda^{-1}$ is larger. In the central part of the Gulf of Alaska $\lambda^{-1}$ ~15 months. Considering a period of 3 yr for the forcing (the Ekman pumping used for the correlations only contains periods longer than 2 yr) the time lag is ~7 months, which is very close to the lag between $W_E$ and OGCM pycnocline depth in the same area.

The pattern of correlations between $h$ and $h_{OGCM}$ obtained with the values of $\lambda^{-1}$ in Figure 7a is shown in Figure 7b. The overall structure is similar to the correlation pattern between $W_E$ and $h_{OGCM}$, but the correlations between $h$ and $h_{OGCM}$ in Figure 7b are higher than the correlations between $W_E$ and $h_{OGCM}$, due to larger level of high-frequency variability in the $W_E$ field with respect to the $h$ field. The correlations in Figure 7b indicate that the simple model gives an excellent fit to the OGCM in the central part of the Gulf of Alaska, and west of ~145°W, with maximum correlations (0.9) at approximately 155°-150°W, 50°-52°N. Correlations decrease onshore, especially towards the Alaska Peninsula and to the south of ~45°N. To the east of Kodiak Island values range from 0.5 south of Queen Charlotte Island to values of 0.7 between Queen Charlotte Island and Kodiak Island, while southwestward of Kodiak Island correlations drop to values close to zero. Thus, the area to the southwest of Kodiak Island is the region where the simple model completely fails to reproduce the OGCM pycnocline depth changes.

To further examine the simple model fit to the OGCM, we compute the difference between pycnocline variability in the OGCM and pycnocline variability produced by equation (2) using the spatially varying $\lambda^{-1}$ that maximizes the correlation. The standard deviation of this residual field, which is a measure of the misfit of the simple model to the OGCM, maximizes in the west-
ern part of the Gulf of Alaska (Figure 8b) where the variance of $h_{OGCM}$ is also very large (Figure 8a). In particular, to the southwest of Kodiak Island the residual field accounts for 80-90% of the variance of $h_{OGCM}$ (Figure 8c). The minimum of the residual standard deviation is found in the central part of the Gulf of Alaska (centered around 150°W, 52°N), where equation (2) produces the best fit to the OGCM. In this area the variance of the residual field accounts for 10-30% of the OGCM pycnocline depth variance (Figure 8c).

5. Rossby wave dynamics

The simple model (2) emphasizes the importance of the local Ekman pumping in driving thermocline variability. Satellite altimetry, on the other hand, has provided evidence of westward propagating SSH anomalies which appear to be associated with first-mode baroclinic Rossby waves (Kelly 1997, Qiu 2002, Fu and Qiu 2002). Which fraction of pycnocline variability can be accounted for by Rossby wave propagation? Could the damping term in equation (2) somehow mimic propagation processes that are not explicitly modelled? To answer these questions we compare the performance of the local Ekman pumping model (2) with that of a Rossby wave model that includes propagation. Fu and Qiu (2002) have examined the evolution of SSH anomalies from the T/P altimeter along 49°N over the period 1993-2000, and have shown that a simple Rossby wave model could reproduce some fundamental aspects of the observed SSH evolution. The pycnocline depth anomalies in the OGCM show an evolution similar to that of the observed SSH anomalies along 49°N over the overlapping period (1993-1997, not shown), and we have chosen to examine the fit of the simple models to the OGCM along 49°N.

Following CL04, we consider a linear, reduced-gravity, quasi-geostrophic model to simulate the upper-ocean variability that includes Rossby wave dynamics. The wavelength of Rossby waves at interannual timescales is typically longer than ~2000km (Chelton and Schlax 1996, Fu and Chelton 2001), while the Rossby radius of deformation is ~20km around 50°N (Chelton et al. 1998). Thus, the long-wave approximation is valid along this latitude. In the long-wave limit the equation describing the evolution of pycnocline depth is:
where \( h_R \) is the pycnocline depth anomaly in the presence of Rossby wave propagation, \( c_R = -\beta R^2 \) is the phase speed of long baroclinic Rossby waves, where \( R \) is the Rossby radius of deformation and \( \beta \) is the latitudinal gradient of the Coriolis parameter, and \( \lambda_1 \) is a Rayleigh friction coefficient. Following Killworth et al. (1997), who took into account the influence of a background mean flow, we have chosen \( c_R = 0.9 \text{ cm s}^{-1} \). To optimize the comparison with the OGCM \( \lambda_1 \) has been chosen to be 1/4 year\(^{-1}\). Equation (4) can be solved by integrating along Rossby wave characteristics in the \( x-t \) plane:

\[
\frac{\partial h_R}{\partial t} + c_R \frac{\partial h_R}{\partial x} = -W_E - \lambda_1 h_R,
\]

Following Killworth et al. (1997), who took into account the influence of a background mean flow, we have chosen \( c_R = 0.9 \text{ cm s}^{-1} \). To optimize the comparison with the OGCM \( \lambda_1 \) has been chosen to be 1/4 year\(^{-1}\). Equation (4) can be solved by integrating along Rossby wave characteristics in the \( x-t \) plane:

\[
h_R(x, t) = h_R(x_E, t - t_E) e^{-\lambda_1 t_E} + \int_{x_E}^{x} \frac{W_E(\xi, t - t_\xi)}{c_R(\xi)} e^{-\lambda_1 t_\xi} d\xi,
\]

where the solution at each point \( x \) and time \( t \) is obtained as the superposition of the disturbances generated east of point \( x \) at previous times. The first term on the right hand side (RHS) of (5) is the contribution of the signals originating at the eastern boundary \( x_E \), and reaching point \( x \) at time \( t \) with a transit time \( t_E = \int_{x_E}^{x} \frac{d\xi}{c_R(\xi)} \). The second term on the RHS of (5) is the contribution of the disturbances generated by the Ekman pumping east of the target point \( x \), with \( t_\xi = \int_{\xi}^{x} \frac{ds}{\xi c_R(s)} \). Both boundary and wind-forced terms decay while propagating counterclockwise with the frictional timescale \( 1/\lambda_1 \).

Shown in Figure 9 is the comparison between the evolution of pycnocline depth anomalies along 49°N from the OGCM and those obtained from the Ekman pumping model (2) and Rossby wave model (4). The pycnocline depth variations in the OGCM (Figure 9a) exhibit a clear change in character after 1976 with positive anomalies west of ~145°W before the shift, and negative afterwards. Pycnocline depth anomalies east of 145°W tend to be of the opposite sign, as implied by Figure 3. Westward propagation is evident for some of the anomalies in Figure 9a, for example the positive anomaly centered at ~155°W in 1965 and the negative anomaly centered around the
same latitude in 1987. However, in most cases pycnocline depth anomalies exhibit a standing character. The evolution produced by the Ekman pumping model (Figure 9b) do not capture the westward propagation seen in the OGCM, but seem to reproduce relatively well the other aspects of the variability. The Rossby wave model, on the other hand, gives a good representation of the westward propagating anomalies, but propagation seems overemphasized by the model compared to the OGCM (Figure 9c).

The success of the Ekman pumping model in explaining pycnocline variability over a large part of the Gulf of Alaska, as described in section 4 could be attributed to the ability of the damping term $-\lambda h$ to mimic unresolved oceanic processes. The results shown in Figure 9 indicate that wave propagation cannot be simulated by a model that does not explicitly include a propagation term. Westward propagating Rossby waves are clearly part of the OGCM dynamics (Figure 9a). However, large-scale anomalies seem to have a standing character that is captured by the local Ekman pumping model. This conclusion is similar to that derived by CL04, based upon the comparison of a quasi-geostrophic model and a local Ekman pumping model with the Topex/Poseidon and Jason-1 altimeter measurements (1993-2003). The comparison of $h_{OGCM}$, $h$, and $h_R$ in Figure 9 also shows that the negative $h_{OGCM}$ anomaly west of $\sim170^\circ W$ in 1974-82 is not captured by neither the Ekman pumping nor the Rossby wave model. The area west of $170^\circ W$ may be affected by advection by the Alaskan Stream, so that some of the anomalies may be due to processes different than local forcing and westward propagation.

6. Coastal propagation of pycnocline depth anomalies

We have seen in section 4 that the agreement between the Ekman pumping model (2) and the OGCM deteriorates in a broad band along the coast, particularly in the western side of the Gulf of Alaska. This band is characterized by the anti-cyclonic circulation of the Alaska gyre, as shown in Figure 10, where the OGCM average velocities above the pycnocline (identified by the 26.4 $\sigma_\theta$ isopycnal) are displayed. The primary currents include a broad northward flowing eastern boundary current, the Alaska Current, which becomes narrower and stronger in the apex of the gyre, and
continues southwestward along the Alaska Peninsula as the Alaskan Stream. Thus, it is conceivable that advection processes may influence the evolution of pycnocline depth anomalies in this area. To examine propagation phenomena along the coastal band we consider a curved path which approximately follows the coast and lies along the axis of the Alaska Current and Alaskan Stream, as shown in Figure 11a. Some points along the path (A, B, C, D, and E) have been highlighted as reference points. Figure 11b shows the time evolution of $h_{OGCM}$ at points A, B, and C, smoothed using a 3-point binomial filter. The OGCM pycnocline variations at point A lag the pycnocline variations at point B by 4 months, with a maximum correlation coefficient of 0.96, and lag the variations at point C by ~6 months, with a maximum correlation coefficient of 0.92. Thus, pycnocline variations along the path are strongly correlated at lags that increase westward, suggesting a counterclockwise propagation of the depth anomalies.

The lag correlations of point A and all the points along the path (Figure 12) shows that the pycnocline depth anomalies are correlated over a very large distance along the coast, and propagate counterclockwise along the path with speeds varying from approximately 16 cm/s at distances larger than 2100 km (approximately east of point C) to ~8 cm/s between 600 km and 2100 km, and decreasing to ~1.3 cm/s at the western end of the path, as indicated by the slope of the dot-dash line in Figure 12. The dot-dash line shows the approximate inclination of the phase lines, and has been computed by fitting a straight line through the points of maximum correlations. Using different points along the path as reference points for the lag correlations leads to different estimates of propagation speeds, especially in the eastern part of the path, approximately east of point C, where propagation appears to be faster when reference points east of point A are considered. In all cases, propagation speeds appear to increase from west to east independently of the reference point chosen.

To account for the propagation of pycnocline depth anomalies, equation (2) has been modified to include a propagation term in the form:

$$\frac{\partial h_p}{\partial t} + v \frac{\partial h_p}{\partial \zeta} = -W_E - R h_p,$$

(6)
where \( h_p \) indicates the isopycnal depth displacement obtained when propagation is included, \( \nu \) is the propagation velocity, \( \zeta \) is the coordinate along the propagation path, and \( R \) is a Rayleigh friction coefficient. We found that the correlations and regressions between \( h_p \) and \( h_{OGCM} \) increase with decreasing values of \( R \). So we have chosen \( R = 1/100 \) year\(^{-1} \). Equation (6) is solved by integrating along the path in Figure 11a:

\[
h_p(\zeta, t) = h_p(\zeta_s, t-t_s) e^{-R t_s} + \int_{\zeta_s}^{\zeta} \frac{W_E(\xi, t-t_s)}{v(\xi)} e^{-R \xi} d\xi.
\]

(7)

The form of the solution is similar to the one obtained for the Rossby wave equation (4), but in this case the solution at each point \( \zeta \) and time \( t \) is obtained as the superposition of the disturbances generated upstream of point \( \zeta \) at previous times. The first term on the right hand side (RHS) of (5), \( h_{PS} \), is the contribution of the signals present at the starting point of the path \( \zeta_s \), and reaching point \( \zeta \) at time \( t \) with a transit time \( t_s = \int_{\zeta_s}^{\zeta} ds \frac{d\xi}{\nu(\xi)} \). The second term on the RHS of (5), \( h_{PE} \), is the contribution of the disturbances generated by the Ekman pumping east of the target point \( \zeta \), with \( t_{\xi} = \int_{\xi_s}^{\xi} ds \frac{d\xi}{\nu(\xi)} \). Both boundary and wind-forced terms decay while propagating counterclockwise with the frictional timescale \( 1/R \).

Using the estimates of propagation speed from the lag-correlations in Figure 12 as a guideline, we have chosen \( \nu(\zeta) \) to increase from 3 cm s\(^{-1} \) at the western end of the path to 7.5 cm s\(^{-1} \) approximately between point A and point C, 10 cm s\(^{-1} \) around point C, and reaching a maximum value of 22 cm s\(^{-1} \) upstream of point C. The fit of equation (6) to the OGCM does not seem to be very sensitive to the values of propagation speed used. Monthly anomalies of pycnocline depth at point E (see Figure 12a) in the OGCM have been used as boundary condition \( h_p(\zeta_s, t) \) in equation (7).

The resulting correlations between \( h_p \) and \( h_{OGCM} \) along the coastal path are shown in Figure 13a, where they are compared with the correlations between \( h_{OGCM} \) and the pycnocline displacements \( h \) obtained with the local Ekman pumping model without propagation. The correlations
between \( h \) and \( h_{OGCM} \) maximize (~0.7) just to the west of point C along the path, but they drop to values close to zero south of point D (4000-5000 km from the western end of the path), and southwest of point B (Kodiak Island). On the contrary, when propagation is included, correlations are ~0.7-0.8 over most of the path, and increase to larger values when approaching the boundary point E. Thus, over most of the path, propagation appears to have a large influence on pycnocline depth evolution. In particular, the pycnocline depth variations to the southwest of Kodiak Island (500-1500 km from the western end) result entirely from the propagation into that area of pycnocline depth anomalies originating further upstream, since the local Ekman pumping would tend to produce anomalies of the opposite sign.

How much of the pycnocline variability along the coastal path is due to the downstream propagation of the displacements at point E (\( h_{ps} \)), and what fraction is forced by the Ekman pumping along the path? The structure of the solution (equation 6) allows to separate the two contributions. Shown by the dotted line in Figure 13a is the correlation between \( h_{ps} \) and \( h_{OGCM} \) along the path, indicating that the downstream propagation of the signals at point E can account for most of the correlations along the path. However, the lower correlations of \( h_{ps} \) and \( h_{OGCM} \) compared to \( h_p \) and \( h_{OGCM} \) downstream of ~3000km from the western end of the path suggest that the Ekman pumping between the distances of 1800km and 3000km may influence the variability. In this area the local Ekman pumping model also yields correlations comparable to those obtained including propagation. The area between 1800 km and 3000 km from the western end of the path is where the largest negative Ekman pumping changes associated with the 1976-77 climate shift are found (Figure 5).

Although the inclusion of propagation considerably improves the performance of the simple model, there are still some discrepancies with the OGCM, especially in the amplitude distributions along the path, as shown by the regression of \( h_p \) and \( h_{OGCM} \) (Figure 13b, solid line): regressions decrease from unity at point E, where the pycnocline depth anomalies coincide with the OGCM values, to ~0.5 at the western end of the path, indicating that the simple model (6) tends to underestimate the amplitude of the pycnocline displacements in the OGCM. When only \( h_{ps} \) is
considered (dotted line in Figure 13b), the amplitude of the OGCM pycnocline variability is further underestimated downstream of point C. The regression between \( h \) and \( h_{OGCM} \) is also shown for comparison in Figure 13b (dot-dash line): apart from the area between points B and C, where the regressions between \( h \) and \( h_{OGCM} \) are comparable to those between \( h_p \) and \( h_{OGCM} \), the local Ekman pumping model evidently underestimates the amplitude of the OGCM pycnocline variability, in particular upstream of point D and downstream of point B, indicating that in those areas propagation is essential. Thus, the correlation and regression analyses suggest that the Ekman pumping forcing can play a role in exciting pycnocline variability in the area around point C, and the wind forced anomalies are then advected downstream.

Are the propagation speeds inferred from the lag-correlation consistent with mean flow advection? The velocities in the OGCM vary strongly with depth, and we need to first determine what is the advective velocity of pycnocline depth. As a conceptual framework we consider a 2-layer quasigeostrophic model with layer thicknesses \( H_1 \) and \( H_2 \) and mean velocities \( U_1 \) and \( U_2 \) for the first and second layers, respectively (Pedlosky 1996). For this system the advective velocity \( U_A \) is:

\[
U_A = \frac{U_2 H_1 + U_1 H_2}{H_1 + H_2}.
\]  

(8)

Thus, if the two layers have equal depths (\( H_1 = H_2 \)), the advective velocity is simply the average velocity \( 0.5(U_1 + U_2) \). If, on the other hand, \( H_1 \ll H_2 \), then \( U_A \sim U_1 \). In winter, the Gulf of Alaska can be viewed as a two layer system, where the two layers are separated by the halocline. The upper layer thickness is \(~100-150\) m, and it is reasonable to assume that \( H_1 \ll H_2 \). Thus, the advective velocity in the model can be estimated by vertically averaging the model velocity from the surface to the depth of the pycnocline. Figure 14 shows the model speed, averaged from the surface to the depth of the 26.4 \( \sigma_\theta \) isopycnal, which we have used as proxy for pycnocline depth, along the coastal path in Figure 11a. The model velocity is not necessarily tangent to the coastal path everywhere, so that the speed in Figure 14 is an estimate of the maximum advective velocity along the path. The upper layer speed increases from 0.5 cm/s at point D to a maximum value of
~6.5 cm/s between point A and Point B (the model’s Alaskan Stream), and then decreases again to ~3 cm/s at the western end of the path. While the upper layer speeds are comparable with the propagation speeds estimated from the lag-correlation analysis to the west of point B, they are much smaller than the propagation speed to the east of point B. Thus, propagation through mean flow advection seems possible to the southwest of Kodiak Island (point B), but further east the model mean velocity is too small to explain the depth anomaly propagation.

Other processes that could be considered include boundary wave propagation. Some observational studies (Enfield and Allen 1980, Chelton and Davis 1982, Qiu 2002) have found evidence of interannual sea level anomalies propagating along the western coast of North America from the Tropics to the Gulf of Alaska. Enfield and Allen (1980) and Chelton and Davis (1982) used sea level data from tide gauges, while Qiu (2002)’s analysis is based on sea surface height (SSH) data from the 8-yr TOPEX/Poseidon mission (1982-2000). Qiu (2002) has shown that the SSH anomalies along the Alaska/Canada coast are partly controlled by equatorial SSH variations through coastally trapped wave dynamics, and are partly forced by the local alongshore winds. The dynamical framework used to demonstrate the importance of the local alongshore winds involves the balance between the onshore (offshore) Ekman transport induced by the alongshore wind stress, and the offshore (onshore) geostrophic transport associated with the alongshore SSH anomaly gradient. Thus, although Qiu (2002)’s results emphasize the importance of both propagation and local wind forcing, his dynamical framework differs from the one adopted in the present study, where Ekman pumping rather than alongshore wind stress is considered.

The observed propagation speeds of the interannual signals estimated through lag-correlations of the tide gauge sea level data range from 40 cm/s (Chelton and Davis 1982) to 87 cm/s (Enfield and Allen 1980). Inviscid boundary waves have an offshore scale of the order of the first baroclinic Rossby radius of deformation, which in the Gulf of Alaska is only ~15-20 km (Chelton et al. 1998), and one may expect a severe distortion of these waves in a model with coarse horizontal resolution. In the presence of lateral viscosity the offshore scale of the Kelvin wave increases, while the alongshore phase speed decreases due to the combined effect of lateral viscosity and
limited model resolution (Hsieh et al. 1983). If the model resolution is fine enough to resolve the broader offshore scale of the viscous wave, then the alongshore phase speed is essentially determined by viscosity, but if the model resolution is still too coarse to resolve the wave structure then the propagation speed is also dependent upon the model resolution.

Given the viscosity used in the model, the offshore scale of the waves may be marginally resolved by the model, but it is unclear how realistic the model’s viscosity is and how distorted boundary waves may be in this numerical simulation. On the other hand, the agreement of the OGCM pycnocline variations with the pycnocline variations observed at station GAK1 suggests that propagation speeds in the model may not be too unrealistic. Further studies are needed to clarify the nature of the propagating signals in the OGCM and the degree of distortion of these signals introduced by the models’s viscosity and resolution.

7. Sverdrup dynamics

A classical model of the mean circulation of the Gulf of Alaska would include a Sverdrup interior with the Alaskan Stream as the western boundary current. Are the changes in the Alaskan Stream that took place after the mid-seventies consistent with Sverdrup dynamics? If the Alaskan Stream does respond to the adjustment of the Alaska Gyre according to Sverdrup dynamics, this could partially explain why its variability is only weekly affected by the local wind stress curl. The time averaged barotropic streamfunction in the model describes the cyclonic circulation of the Alaska Gyre (Figure 15a). The mean transport of the model Alaskan Stream is ~7-8 Sv, a value that is much smaller than some observational estimates. Geostrophic transports relative to 1500 dbar are typically 12-18 Sv (Musgrave et al. 1992). However, intermittent surveys of the Alaskan Stream yielded a broad range of transport estimates, with values as low as 2 Sv and as large as 20 Sv (Reed et al. 1980, Royer 1981). The model underestimate of the Alaskan Stream transport is likely the result of the model’s coarse horizontal resolution and relatively high viscosity. We believe that in spite of the OGCM inaccuracies in the representation of the Alaska Gyre circulation our results would remain qualitatively correct.
The barotropic streamfunction difference between 1977-97 vs. 1960-75 (Figure 15b) shows a slight weakening of the circulation in the eastern part of the basin, and a strengthening of the circulation to the southwest of Kodiak Island. The intensification of the Alaskan Stream after the mid-seventies is consistent with the more pronounced doming of the isopycnals after the mid-seventies (Figure 2), and resulting larger zonal density gradients (Figure 3).

The Sverdrup’s (1947) balance is derived from the vertical integral of the vorticity equation

\[ \beta v = f \frac{\partial w}{\partial z}, \]  

where \( v \) is the meridional velocity, \( w \) is the vertical velocity, and \( z \) the vertical coordinate, stating that the changing in potential vorticity associated with the vertical stretching of the water column must be balanced by the changes in planetary vorticity due to meridional motions. Integrating (8) from a level of no-motion to the base of the Ekman layer, we obtain the classical form of the Sverdrup balance:

\[ \beta V_g = f W_E, \]  

where \( V_g \) is the vertically integrated geostrophic meridional velocity. The condition of non-divergence for the vertically integrated geostrophic flow allows to introduce a streamfunction \( \psi_g \) such that

\[ V_g = \frac{\partial \psi_g}{\partial x}. \]  

Using the boundary condition that \( \psi_g(x_E) = 0 \) at the eastern boundary \( x_E \) of the ocean basin, the geostrophic streamfunction at longitude \( x \) along a given latitude can be computed by integrating westward from the eastern boundary:

\[ \psi_g(x) = -\frac{f}{\beta} \int_{x_E}^{x} W_E dx'. \]  

The changes in barotropic streamfunction associated with the 1976-77 climate shift, as predicted by equation (11), have been computed using the Ekman pumping difference in Figure 5. The result (Figure 15c) shows significant differences with the streamfunction changes found in the OGCM. The circulation changes implied by the Sverdrup balance include an intensification of
the Alaskan Stream only south of 54°N, while north of this latitude the flow weakens appreciably, with transport changes as large as 2.5Sv. Thus, the Sverdrup balance does not seem to provide a good description of the wind forced circulation changes in the Gulf of Alaska after the mid-seventies. The Sverdrup balance is achieved through Rossby wave propagation. As seen in section 5, Rossby wave propagation can account for only a part of the pycnocline variability in the OGCM. On the other hand, propagation of pycnocline depth anomalies along the coast seems to have a significant impact on the pycnocline changes along the Alaska Peninsula (section 6), and may be responsible for some of the changes in the density structure and related circulation changes.

8. Conclusions

In this paper we have examined the influence of local Ekman pumping and propagation processes upon pycnocline depth variability in the Gulf of Alaska, using the output from an ocean general circulation model forced with observed surface fields over the period 1958-1997. The evolution of the OGCM pycnocline depth has been compared with that observed at two locations in the Gulf of Alaska, where long-term observations of the upper-ocean density structure are available. The two observational stations are Ocean Weather Station P, located in the central part of the Gulf of Alaska (50°N, 145°W), and station GAK1, located at the mouth of Resurrection Bay along the Seward line (59°N, 149°W). The winter pycnocline depth changes in the OGCM agree relatively well with those observed at both interannual and longer timescales, especially during periods in which the density of observations was higher. The low-frequency evolution of the pycnocline depth is dominated by the 1976-77 climate regime shift. After the mid-seventies, the model shows a deepening of the pycnocline in a broad band following the coast, and shoaling of the pycnocline in the central part of the Gulf of Alaska. These changes are consistent with the low-frequency evolution of the pycnocline depth at station GAK1, where the pycnocline exhibits a deepening trend, and at OWS P, where a shoaling trend is observed.

To understand what fraction of the OGCM pycnocline depth variability can be driven by the local Ekman pumping we have introduced a simple model for the evolution of pycnocline depth,
where the forcing is supplied by the local Ekman pumping and dissipative processes are modeled with a linear damping term. This model is very similar to the one used by Lagerloef (1995) and CL02. Solutions to the Ekman pumping model have been computed for different values of the damping timescale. Correlations between pycnocline depth variations from the simple model and from the OGCM maximize toward the center of the Alaska Gyre, where the solution of the simple model accounts for a large fraction of the OGCM pycnocline depth variability. Correlations decrease along a broad coastal band, and in particular to the southwest of Kodiak Island, where they drop to values very close to zero. In this area the pycnocline deepens, while the Ekman pumping changes are indicative of enhanced upwelling. The damping timescales that yield the maximum correlations between the OGCM and the simple model are characterized by a spatial pattern with values ranging from 14 - 18 months in the central part of the Gulf of Alaska to values as large as 90 months at some locations along the coast. The damping timescales in the central part of the Gulf of Alaska, where correlations are higher, are very similar to those estimated by Lagerloef (1995) and CL02 using statistical methods.

Westward propagating Rossby waves can be detected in the evolution of the model pycnocline. Propagating features cannot be reproduced by the local Ekman pumping model, but can be modeled with the explicit inclusion of a zonal propagation term. However, large-scale anomalies have a predominantly standing character which is well-captured by the Ekman pumping model.

Pycnocline depth anomalies exhibits a counterclockwise propagation along the coastal band, clearly indicating that Ekman pumping alone cannot account for the pycnocline variability in that area. A propagation term has been added to the equation governing the pycnocline depth evolution, using a propagation velocity that has been estimated from the OGCM by lag-correlations. The modified equation has been solved along a path following the coast, and the results indicate that the inclusion of propagation considerably improves the agreement with the OGCM along the path. An open question remains concerning the physical nature of the propagation velocity. Along the western half of the path pycnocline depth anomalies appear to propagate with a speed that is very similar to the OGCM upper layer mean speed, so that propagation could be consistent with
mean flow advection. However, the propagation speed increases eastward to values that are much larger than the upper layer mean speed in the OGCM. Thus, in the eastern part of the path wave processes appear to be a more likely cause of pycnocline depth anomalies propagation, but coastal waves may be distorted in the model due to excessive viscosity and coarse resolution.

The large changes in the upper-ocean density structure and circulation that took place after the mid-seventies do not seem to result from the dynamical adjustment of the Alaska gyre according to Sverdrup’s theory. The latter would imply a weakening of the circulation north of ~54°N, while the circulation changes in the model exhibit an intensification of the Alaskan Stream starting upstream of Kodiak Island. However, the intensification of the Alaskan Stream is consistent with the changes in the upper-ocean density structure associated with the 1976-77 climate shift. The deepening of the pycnocline along the Alaskan Stream results primarily from the propagation of pycnocline depth anomalies along the coast, suggesting that in Gulf of Alaska the coastal waveguide may supply a faster route for the adjustment of the Alaskan Stream than the interior Rossby waves.

The upper ocean in the Gulf of Alaska is characterized by a fresh and well mixed surface layer separated from the lower layer by a halocline. Since at high latitudes density is largely controlled by salinity, the pycnocline approximately coincides with the halocline, and also defines the depth of the winter mixed layer. Thus, the low-frequency variability of pycnocline depth is also an indication of variability in mixed layer depth, a quantity that plays a fundamental role in biological productivity (Gargett 1997). It is plausible that the changes in pycnocline depth described in this study may help understand the widespread environmental changes that have accompanied the 1976-77 climate regime shift, in particular the decline of steller sea lion populations in the western Gulf of Alaska (Trites and Donnelly 2003). Future studies are necessary to elucidate the links between the physical changes in the upper-ocean thermal structure and biological activity.

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References


Figure Captions

Figure 1. Evolution of potential temperature (a), salinity (b) and potential density (c) at the location of Ocean Weather Station Papa (50°N, 145°W) from July 1995 to June 1996 based upon the output of the NCAR OGCM. Contour intervals are 0.5°C for temperature, 0.1 ppt for salinity, and 0.2 σθ units for density. The thicker contour in c) indicates the 26.4 σθ isopycnal.

Figure 2. Zonal section of mean potential density along 53.8°N for the periods 1958-75 (solid contours) and 1977-97 (dot-dashed contours). Contour interval is 0.4 σθ units. The 26.4 σθ isopycnal, which is used as proxy for pycnocline depth, is highlighted with a thick contour for both periods.

Figure 3. Pycnocline depth changes associated with the 1976-77 climate regime shift from the output of the NCAR OGCM. The depth of the 26.4 σθ isopycnal is used as proxy for pycnocline depth. The positions of Ocean Weather Station P (P, 145W, 50N), GAK1 (G, 149W, 59N), and the location of Kodiak Island (K), Seward (S), Queen Charlotte Island (Q), the Alaska Peninsula, and the Aleutian Islands are indicated for reference.

Figure 4. Pycnocline depth variations (in meters) at OWS P (top), and station GAK1 (bottom). Positive anomalies indicate deeper pycnocline, while negative anomalies correspond to a shallower pycnocline. The dots connected by the dashed line are the winter averages of observed values at the two stations (usually a few values for each winter), while the solid line shows the winter pycnocline depth anomalies from the NCAR OGCM, based on monthly averages. Correlations coefficients between the observed and modeled winter values are 0.7 and 0.77 at Papa and GAK1, respectively. The observed time series at Papa was kindly provided by Dr. P. Cummins.

Figure 5. Winter Ekman pumping changes associated with the 1976-77 climate regime shift. Contour interval is 0.1 $10^{-4}$ cm/s. Solid contours indicate positive values, while dashed contours are used for negative values. Anomalies larger than 0.3 $10^{-4}$ cm/s are dark shaded, while values
smaller than -0.3 \times 10^{-4} \text{ cm/s} are light shaded. The mean Ekman pumping is positive over the Gulf of Alaska. Negative Ekman pumping anomalies imply reduced upwelling and vice versa.

**Figure 6.** a) Maximum lag-correlations between low-pass filtered Ekman pumping (periods shorter than 2 yr are removed) and pycnocline depth anomalies. Pycnocline depth anomalies have been multiplied by -1 to achieve positive correlations (a positive depth anomaly indicates deeper pycnocline, and should be associated with negative $W_E$ anomalies, and vice versa). Values larger than 0.5 are light-shaded, while values larger than 0.7 are dark-shaded. b) Lag (in months) between $W_E$ and $h_{OGCM}$ at which the maximum correlations in a) are achieved. Positive lags indicate that $W_E$ leads the changes in pycnocline depth.

**Figure 7.** a) Spatial distribution of the damping timescales $\lambda^{-1}$ (in months) which maximize the correlation between pycnocline depth anomalies produced by the Ekman pumping model (2) and the pycnocline depth anomalies in the OGCM. b) Spatial distribution of the correlation between $h$ and $h_{OGCM}$ obtained using the values of $\lambda$ in a). Light shading is for values larger than 0.7, intermediate shading is for values larger than 0.8, and dark shading is for values larger than 0.9.

**Figure 8.** a) Standard deviation of pycnocline depth in the OGCM. Contour interval is 5 m. Values larger than 10 m are shaded. b) Standard deviation of the residual field computed as the difference between pycnocline depth anomalies in the OGCM and pycnocline depth anomalies from the Ekman pumping model (2). Contour interval is 5 m. Values larger than 10 m are shaded. c) Percentage of the OGCM pycnocline depth variance accounted for by the residual field. Values larger than 60% are light-shaded, and values larger than 70% are dark-shaded.

**Figure 9.** Comparison between the evolution of pycnocline depth along 49°N in the OGCM (a), as computed from the Ekman pumping model (b), and the one obtained from the Rossby wave model (c). The solution for the Ekman pumping model was obtained by using the optimal values.
of \( \lambda \) in Figure 7a along 49°N. Dashed contours and blue shading are for negative pycnocline depth anomalies (shallower pycnocline), while solid contours and orange shading are for positive anomalies (deeper pycnocline). Contour interval is 10 m in all three panels.

**Figure 10.** Average upper-layer velocities in the OGCM. The upper-layer is defined as the layer extending from the surface to the depth of the 26.4 \( \sigma_\theta \) isopycnal, which is used as proxy for pycnocline depth. Units are cm s\(^{-1}\).

**Figure 11.** a) Coastal path along which lag-correlations of pycnocline depth anomalies in the OGCM are considered. The points ‘A’, ‘B’, ‘C’, ‘D’, and ‘E’ are highlighted for reference. b) time series of \( h_{OGCM} \) at point A (solid line), B (dot-dash line), and C (dotted line). The time series have been smoothed with a 3-point binomial filter. The pycnocline variability at point A lags the pycnocline variability at point B by \( \sim \)4 months (correlation coefficient = 0.96), and the pycnocline variability at point C by \( \sim \)6 months (correlation coefficient = 0.92).

**Figure 12.** Lag-correlations between pycnocline depth variations at point A in Figure 8a and all the points along the coastal path. Values larger than 0.7 are light-shaded, and values larger than 0.9 are dark-shaded. The dot-dashed line indicates the approximate slope of the phase lines in three different areas along the path, corresponding to propagation speeds of \( \sim \)1.3 cm/s, 8.6 cm/s, and 16 cm/s from west to east. The lags yielding the maximum correlation at each distance have been used as a guideline to draw the dot-dash lines. Distance is computed starting at the westernmost point. To associate the distance with specific locations along the path, the points A, B, C, D, and E in Figure 8a are indicated along the top axis.

**Figure 13.** (Top) The solid line shows the correlation between isopycnal depth displacements obtained from the Ekman pumping model with propagation (\( h_p \)) and the isopycnal depth displacements in the OGCM (\( h_{OGCM} \)) along the coastal path in Figure 8a. The dot-dashed line shows the
correlation between the isopycnal depth variations from the Ekman pumping model without propagation \((h)\) and \(h_{OGCM}\) along the path. (Bottom) Same as top, but regressions are displayed instead of correlations.

**Figure 14.** Winter (December - April) upper ocean speed as a function of distance along the coastal path in Figure 8a in the OGCM. The upper ocean speed has been estimated by vertically integrating the model speed from the surface to the depth of the 24.6 \(\sigma_\theta\) isopycnal, assumed as proxy for pycnocline depth. The points A, B, C, and D shown in Figure 8a are indicated along the top axis for orientation purposes.

**Figure 15.** a) Mean model barotropic streamfunction over the Gulf of Alaska, describing the cyclonic circulation of the model Alaska Gyre. b) Difference of barotropic streamfunction between 1977-97 and 1960-75. c) Difference in barotropic streamfunction implied by Sverdrup dynamics, computed as the zonal integral of the Ekman pumping pattern in Figure 5. Contour intervals are 2.5 \(Sv\) in a), and 0.5 \(Sv\) in b) and c). Solid contours are for positive values, and dashed contours are for negative values.
Figure 1. Evolution of potential temperature (a), salinity (b) and potential density (c) at the location of Ocean Weather Station Papa (50°N, 145°W) from July 1995 to June 1996 based upon the output of the NCAR OGCM. Contour intervals are 0.5°C for temperature, 0.1 ppt for salinity, and 0.2 $\sigma_{\theta}$ units for density. The thicker contour in c) indicates the 26.4 $\sigma_{\theta}$ isopycnal.
Figure 2. Zonal section of mean potential density along 53.8°N for the periods 1958-75 (solid contours) and 1977-97 (dot-dashed contours). Contour interval is 0.4 $\sigma_\theta$ units. The 26.4 $\sigma_\theta$ isopycnal, which is used as proxy for pycnocline depth, is highlighted with a thick contour for both periods.
Figure 3. Pycnocline depth changes associated with the 1976-77 climate regime shift from the output of the NCAR OGCM. The depth of the 26.4 $\sigma_{\theta}$ isopycnal is used as proxy for pycnocline depth. The positions of Ocean Weather Station P (P, 145°W, 50°N), GAK1 (G, 149°W, 59°N), and the location of Kodiak Island (K), Seward (S), Queen Charlotte Island (Q), the Alaska Peninsula, and the Aleutian Islands are indicated for reference.
Figure 4. Pycnocline depth variations (in meters) at OWS P (top), and station GAK1 (bottom). Positive anomalies indicate deeper pycnocline, while negative anomalies correspond to a shallower pycnocline. The dots connected by the dashed line are the winter averages of observed values at the two stations (usually a few values for each winter), while the solid line shows the winter pycnocline depth anomalies from the NCAR OGCM, based on monthly averages. Correlations coefficients between the observed and modeled winter values are 0.7 and 0.77 at Papa and GAK1, respectively. The observed time series at Papa was kindly provided by Dr. P. Cummins.
Figure 5. Winter Ekman pumping changes associated with the 1976-77 climate regime shift. Contour interval is $0.1 \times 10^{-4}$ cm/s. Solid contours indicate positive values, while dashed contours are used for negative values. Anomalies larger than $0.3 \times 10^{-4}$ cm/s are dark shaded, while values smaller than $-0.3 \times 10^{-4}$ cm/s are light shaded. The mean Ekman pumping is positive over the Gulf of Alaska. Negative Ekman pumping anomalies imply reduced upwelling and vice versa.
Figure 6. a) Maximum lag-correlations between low-pass filtered Ekman pumping (periods shorter than 2 yr are removed) and pycnocline depth anomalies. Pycnocline depth anomalies have been multiplied by -1 to achieve positive correlations (a positive depth anomaly indicates deeper pycnocline, and should be associated with negative $W_E$ anomalies, and vice versa). Values larger than 0.5 are light-shaded, while values larger than 0.7 are dark-shaded. b) Lag (in months) between $W_E$ and $h_{OGCM}$ at which the maximum correlations in a) are achieved. Positive lags indicate that $W_E$ leads the changes in pycnocline depth.
Figure 7. a) Spatial distribution of the damping timescales $\lambda^{-1}$ (in months) which maximize the correlation between pycnocline depth anomalies produced by the Ekman pumping model (2) and the pycnocline depth anomalies in the OGCM. b) Spatial distribution of the correlation between $h$ and $h_{OGCM}$ obtained using the values of $\lambda$ in a). Light shading is for values larger than 0.7, intermediate shading is for values larger than 0.8, and dark shading is for values larger than 0.9.
Figure 8. a) Standard deviation of pycnocline depth in the OGCM. Contour interval is 5 m. Values larger than 10 m are shaded. b) Standard deviation of the residual field computed as the difference between pycnocline depth anomalies in the OGCM and pycnocline depth anomalies from the Ekman pumping model (2). Contour interval is 5 m. Values larger than 10 m are shaded. c) Percentage of the OGCM pycnocline depth variance accounted for by the residual field. Values larger than 60% are light-shaded, and values larger than 70% are dark-shaded.
Figure 9. Comparison between the evolution of pycnocline depth along 49°N in the OGCM (a), as computed from the Ekman pumping model (b), and the one obtained from the Rossby wave model (c). The solution for the Ekman pumping model was obtained by using the optimal values of $\lambda$ in Figure 7a along 49°N. Dashed contours and blue shading are for negative pycnocline depth anomalies (shallower pycnocline), while solid contours and orange shading are for positive anomalies (deeper pycnocline). Contour interval is 10 m in all three panels.
Figure 10. Average upper-layer velocities in the OGCM. The upper-layer is defined as the layer extending from the surface to the depth of the 26.4 $\sigma_\theta$ isopycnal, which is used as proxy for pycnocline depth. Units are cm s$^{-1}$. 
Figure 11. a) Coastal path along which lag-correlations of pycnocline depth anomalies in the OGCM are considered. The points 'A', 'B', 'C', 'D', and 'E' are highlighted for reference. b) Time series of $h_{OGCM}$ at point A (solid line), B (dot-dash line), and C (dotted line). The time series have been smoothed with a 3-point binomial filter. The pycnocline variability at point A lags the pycnocline variability at point B by ~4 months (correlation coefficient = 0.96), and the pycnocline variability at point C by ~6 months (correlation coefficient = 0.92).
Figure 12. Lag-correlations between pycnocline depth variations at point A in Figure 8a and all the points along the coastal path. Values larger than 0.7 are light-shaded, and values larger than 0.9 are dark-shaded. The dot-dashed line indicates the approximate slope of the phase lines in three different areas along the path, corresponding to propagation speeds of ~1.3 cm/s, 8.6 cm/s, and 16 cm/s from west to east. The lags yielding the maximum correlation at each distance have been used as a guideline to draw the dot-dash lines. Distance is computed starting at the westernmost point. To associate the distance with specific locations along the path, the points A, B, C, D, and E in Figure 8a are indicated along the top axis.
Figure 13. a) The solid line shows the correlation between isopycnal depth displacements obtained from the Ekman pumping model with propagation ($h_p^*$) and the isopycnal depth displacements in the OGCM ($h_{OGCM}$) along the coastal path in Figure 8a. The dot-dashed line shows the correlation between the isopycnal depth variations from the Ekman pumping model without propagation ($h$) and $h_{OGCM}$ along the path. The dotted line shows the correlation between the pycnocline depth anomalies due to the propagation of the signal at the starting point E ($h_{PS}^*$) and the pycnocline depth anomalies in the OGCM. b) Same as top, but regressions are displayed instead of correlations.
Figure 14. Winter (December - April) upper ocean speed as a function of distance along the coastal path in Figure 8a in the OGCM. The upper ocean speed has been estimated by vertically integrating the model speed from the surface to the depth of the $26.4 \sigma_t$ isopycnal, assumed as proxy for pycnocline depth. The points A, B, C, and D shown in Figure 8a are indicated along the top axis for orientation purposes.
Figure 15. a) Mean model barotropic streamfunction over the Gulf of Alaska, describing the cyclonic circulation of the model Alaska Gyre. b) Difference of barotropic streamfunction between 1977-97 and 1960-75. c) Difference in barotropic streamfunction implied by Sverdrup dynamics, computed as the zonal integral of the Ekman pumping pattern in Figure 5. Contour intervals are 2.5 $Sv$ in a), and 0.5 $Sv$ in b) and c). Solid contours are for positive values, and dashed contours are for negative values.