Decadal Variability in the Northeast Pacific in a Physical-Ecosystem Model: The Role of Mixed Layer Depth and Trophic Interactions

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Abstract

A basin-wide interdecadal change in both the physical state and the ecology of the North Pacific occurred near the end of 1976. Here we use a physical-ecosystem model to examine whether changes in the physical environment associated with the 1976-77 transition influenced the lower trophic levels of the food web and if so by what means. The physical component is an ocean general circulation model, while the biological component contains ten compartments: two phytoplankton, two zooplankton, two detritus pools, nitrate, ammonium, silicate and TCO$_2$. The model is forced with observed atmospheric fields during 1960-1999. During spring, there is a ~40% reduction in plankton biomass in all four plankton groups in spring during 1977-88 relative to 1970-76 in the central Gulf of Alaska (GOA). The epoch difference in plankton appears to be controlled by the mixed layer depth. Enhanced Ekman pumping after 1976 caused the halocline to shoal, and thus the mixed layer depth did not penetrate as deep in the central GOA during late winter. As a result, more phytoplankton remained in the euphotic zone and phytoplankton biomass began to increase earlier in the year after the 1976 transition. Zooplankton biomass also increased but then grazing pressure led to a strong decrease in phytoplankton by April followed by a drop in zooplankton by May: essentially the mean seasonal cycle of plankton biomass was shifted earlier in the year. Finally, plankton concentrations rebounded slightly after 1976, leading to a small increase in zooplankton biomass during summer in 1977-88 relative to 1970-76.
1. Introduction

Studies conducted over the past 15-20 years have provided a growing body of evidence for decadal climate variability across the Pacific Basin [e.g. Trenberth, 1990, Graham, 1994, Zhang et al., 1997; Deser et al., 2004]. The most prominent mode of decadal variability in the North Pacific was termed the Pacific Decadal Oscillation (PDO) by Mantua et al. [1997] based on transitions between relatively stable states of the leading pattern of sea surface temperature (SST) anomalies. The 1976-77 transition or “regime shift” was especially pronounced, with an increase in the strength of the atmospheric circulation over the North Pacific that resulted in basin-wide changes in ocean temperatures, currents and mixed layer depth [e.g. Miller et al., 1994; Trenberth and Hurrell, 1994; Polovina et al., 1995; Deser et al., 1996, 1999]. These climatic changes had a pervasive effect on marine ecosystems from phytoplankton to the top trophic levels [e.g. Mantua et al., 1997; Hare and Mantua, 2000; Benson and Trites, 2002].

While the dynamics underlying Pacific decadal variability have not been fully resolved, changes in the physical state of the ocean clearly influence the primary and secondary production in the North Pacific. For example, total chlorophyll a, a proxy for phytoplankton biomass, nearly doubled in the central North Pacific from 1968 to 1985 [Venrick et. al., 1987]. Brodeur and Ware (1992), Brodeur et al. [1996] and Rand and Hinch [1998] found that the zooplankton biomass in summer was relatively high in the central portion of the Gulf of Alaska (GOA) during 1956-1962, while in 1980-1989 the biomass was greatest along the edges of the Alaskan Gyre. Overall, the zooplankton biomass nearly doubled over the northeast Pacific in the 1980s relative to the late 1950s.
and early 1960s. McFarlane and Beamish [1992] also found that the copepod biomass increased after the winter of 1976-77 in the Gulf of Alaska. In contrast, Sugimoto and Tadokoro [1997], found a decrease in both phytoplankton and zooplankton biomass in the northeast Pacific during the summers of 1980-1994 relative to those in 1960-1975. Some of the differences between these estimates of decadal variations may be due to the large interannual and spatial variability of zooplankton biomass, especially given the limited number of measurements.

Several physical/chemical factors may influence primary productivity in the Pacific on interannual to decadal time scales, including temperature, sunlight, macronutrients such as nitrogen and silica, and micronutrients, especially iron [see reviews by Francis et al., 1998; Miller et al., 2004]. Vertical mixing, and the mixed layer depth (MLD) in particular, are critical in linking the physical/chemical and biological processes [e.g. Mann, 1993; Steele and Henderson, 1993; Gargett, 1997]. If the mixed layer is too deep, phytoplankton will be transported beneath the euphotic zone inhibiting their growth due to the absence of light, but if it is too shallow, not enough nutrients will be mixed into the surface layer from below to sustain photosynthesis. In addition, if the MLD is sufficiently shallow in late winter for photosynthesis to occur and the resulting phytoplankton biomass is adequate to sustain zooplankton populations, then when the mixed layer shoals in spring, grazing by zooplankton can limit algal growth rates [e.g. Parslow, 1985; Frost, 1991; Fasham, 1995].

Several studies have emphasized the impact of changes in MLD during 1976-77 on the lower trophic levels of the North Pacific ecosystems [Venrick et al., 1987; Polovina et al., 1994]. Brodeur and Ware [1992] hypothesized that changes in wintertime MLD over
the northeast Pacific could impact the micronutrient supply and/or change the rate of primary productivity and the efficiency of zooplankton grazing in spring. Polovina et al. [1995] indicated that shoaling of the mixed layer in winter post 1976-77 could significantly enhance productivity in the Gulf of Alaska by increasing the light available for photosynthesis based on MLD estimated from temperature profiles and a plankton population dynamics model. However, the MLD is influenced by salinity in subarctic waters and it is unclear whether there is sufficient variability in the late winter MLD in the northeast Pacific to significantly impact primary and secondary production [McClain et al., 1996].

Mixed layer depth is regulated by wind stirring, buoyancy forcing via surface heat and freshwater fluxes, and the density jump at the base of the mixed layer [Alexander et al., 2000]. During late winter in the subarctic North Pacific, the mixed layer extends down to the upper portion of the halocline, located between depths of approximately 70 m to 120 m [Fig. 3; Roden, 1964; Freeland et al., 1997; de Boyer Montégut et al., 2004]. Thus, low-frequency changes in the Ekman pumping in the Gulf of Alaska, which vertically displaces the halocline, may impact the wintertime MLD. Indeed, after the mid-1970s, the pycnocline was shallower in the central part of the Gulf of Alaska and deeper in a broad band along the coast, primarily due to the local response to Ekman pumping [Cummins and Lagerloef, 2002; Capotondi et al., 2005].

Decadal changes in the physical state of the North Pacific Ocean during the 1976 transition and their impact on the phytoplankton and zooplankton were simulated by Haigh et al. [2001] and Chai et al. [2003] using an ecosystem model embedded in an ocean general circulation model (OGCM). Haigh et al. performed simulations for the
years 1952-1975 and for 1977-1989, using the mean atmospheric forcing over those two periods. The model reproduced the observed increase in phytoplankton in the central Pacific in summer and a southward displacement of the subtropical chlorophyll front in all seasons in the later period relative to the earlier one. The model also indicated an increase in zooplankton over the eastern subarctic Pacific in summer after 1976, but the magnitude and pattern of the increase differed from the observed values in Brodeur and Ware [1992]. Chai et al. [2003] used time-dependent forcing and substantially different physical and biological components than Haigh et al. [2001] to investigate variability in the subtropical chlorophyll front. They found that the front expanded and extended farther south after the 1976-77 shift, in agreement with Haigh et al.. In the present study, we use the model simulation described by Chai et. al. [2003] to investigate the physical and biological changes in the northeast Pacific Ocean, especially those which occurred across the 1976-77 transition.

2. The Model Simulation

The physical component of the model, which simulates temperature, salinity and currents, is the National Center for Atmospheric Research (NCAR) ocean model (NCOM), described by Large et al. [1997] and Gent et al. [1998]. 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 non-local planetary boundary layer parameterization [Large et al., 1994]. The advection of tracers, including temperature, salinity and the biogeochemical components, is calculated using a third-order differencing scheme. The
background horizontal and vertical mixing coefficients for all tracers are $2.0 \times 10^6$ cm$^{-2}$ sec$^{-1}$ and 0.1 cm$^{-2}$ sec$^{-1}$, respectively. The version of the model used here, described by Li et al. [2001], covers the Pacific from 45°S to 65°N and from 100°E to 70°W with realistic geometry and bathymetry. There is no flow through the boundaries and a sponge layer, where the model’s temperature, salinity, nitrate and silicate are relaxed towards observations, is applied within 10° of the meridional boundaries. The longitudinal resolution is a uniform 2°. The latitudinal resolution is 0.5° within 10° of the equator, gradually increases to 2° within 20° of the equator, and is fixed at 2° north of 20°N and south of 20°S. There are 40 vertical levels, with 23 in the upper 400 m. The model’s relatively coarse horizontal and vertical resolution allows us to examine basin-scale variability over an extended period of time, but is not sufficient to resolve eddies or coastal processes.

The biological model, developed by Chai et al. [2002], consists of ten compartments with two classes of phytoplankton (P1, P2) and zooplankton (Z1, Z2), two forms of dissolved inorganic nitrogen: nitrate (NO$_3$) and ammonium (NH$_4$), detrital nitrogen (DN), silicate (Si(OH)$_4$), detrital silicon (DSi), and total carbon dioxide (TCO$_2$). Small phytoplankton (P1) have variable growth rates, that depends on temperature, nitrogen and light. Their biomass is regulated by micrograzers (Z1), while their net productivity is largely remineralized [Landry et al., 1997]. The large phytoplankton class (P2 > 10 µm) represents the diatom functional group, which can grow rapidly under optimal nutrient conditions [Coale et al., 1996]. Iron and its limitation on phytoplankton growth is not modeled directly but is treated implicitly via two parameters: the slope of the photosynthetic rate over irradiance at low irradiance and the maximum phytoplankton
growth rate [Chai et al., 2002; Chai et al., 2007]. The parameter values, which are constant in space and time, were derived from the limited number of available lab and field experiments, and model simulations of iron addition experiments in the equatorial Pacific (Chai et al., 2007). The micrograzers have growth rates similar to P1 and grazing rates (G1) that depend on the density of both P1 and Z1 [Landry et al., 1997]. The mesozooplankton (Z2) graze on P2 and DN and prey on Z1 and have a feeding threshold based on conventional grazing dynamics [Frost and Frazen, 1992]. The loss of Z2 from the system, primarily due to predation from higher trophic levels, is represented by a quadratic expression. Sinking particulate organic matter is converted to inorganic nutrients via regeneration, similar to the process described by Chai et al. [1996]. Nitrogen is the “currency” in the ecosystem model, i.e. plankton biomass and detritus pools are in units of millimoles of nitrogen per cubic meter (mmol N m\(^{-3}\)). A detailed discussion of the model equations and parameter values is given in Chai et al. [2002].

The model’s temperature, salinity and nutrients were initialized using climatological values from the National Ocean Data Center (NODC), while the biological components were assigned a value of 0.025 mmol m\(^{-3}\) at the surface, decreasing exponentially with a scale length of 120 m – the average depth of the euphotic zone. The full model was then integrated for 10 years with climatological forcing, during which it reached a stable annual cycle in the upper ocean Chai et al. [2003]. During this spin-up period and in the subsequent model experiment the physical and ecosystem components were integrated synchronously.

In the simulation examined here, the model is forced with observed atmospheric fields over the period 1955-1999 and the output is archived monthly during 1960-1999,
allowing for a five-year spin-up period. The surface fluxes of momentum, heat, fresh water, and insolation used to drive the model were derived from monthly mean values obtained from the Comprehensive Ocean Atmosphere Data Set [COADS; da Silva et al., 1994] over the period 1955-1993. From 1993-1999 the surface fluxes were obtained from the National Center for Environmental Prediction (NCEP) reanalysis [Kalnay et al., 1996; Kistler et al., 2001]. The heat flux includes shortwave and longwave radiation and the sensible and latent heat flux. The sensible (latent) fluxes are computed using the observed air temperature (specific humidity) and the model’s SST. The model does not include the diurnal cycle, so the daily averaged insolation is used for calculating photosynthesis. The model is forced by the observed difference between precipitation and evaporation derived from COADS. Since precipitation is not well measured over the ocean, the salinity of the surface layer is also relaxed to the observed climatological monthly mean values [Levitus et al., 1994] using a 30-day time scale. While this damping greatly reduces the surface salinity variability, the deeper ocean, including the halocline, is relatively unconstrained. Li et al. [2001] and Chai et al. [2003] provide more detailed descriptions of the procedures used to initialize and force the model and its fidelity in simulating the physical and biological state of the Pacific Ocean.

3. Results

3.1 SST Changes

We will mainly focus on the years 1970-76 and 1977-88 since they were dominated by opposite phases of the PDO [Bond et al., 2003] and because the fields used to force the model during these years were all derived from COADS (NCEP reanalysis is used to
drive the model after 1993). The epoch difference (Δ, defined as 1977-1988 minus 1970-1976 from hereon) in the observed and simulated North Pacific SST (°C) during February-March-April (FMA) is shown in Fig. 1. In both observations and the OGCM, negative ΔSST in the central and western Pacific between approximately 25°N-45°N is ringed by positive ΔSST over the remainder of the North Pacific, indicative of a change to the positive phase of the PDO after 1976. The model also reproduces several of the finer-scale features of the ΔSST field, with negative centers at 30°N, 155°W and 40°N, 175°E, and positive centers in the vicinity of the Alaskan Peninsula and Baja California. The latter is part of a broad swath the extends over much of the tropical Pacific suggesting that the PDO is part of a larger pattern that includes the tropics as well as the North Pacific, consistent with the findings of Zhang et al. [1997], Newman et al. [2003] and Deser et al. [2004]. The close correspondence between the observed and simulated ΔSST fields provides some confidence in the GCM’s ability to simulate decadal changes in the northeast Pacific, although using the observed air temperature to compute the surface fluxes partially constrains the SSTs to track observations [e.g. Seager et al., 1995].

3.2 Physical – Biological linkages

A number of studies, including Brodeur and Ware [1992], Polovina et al. [1995] and Freeland et al. [1997], indicated that surface mixing in winter has a strong impact on the Gulf of Alaska ecosystem over the course of the seasonal cycle. The ΔMLD (m) during February-April (FMA) is shown for the GOA (contours) in Fig. 2, where the MLD is defined as the depth where the potential density is 0.125 kg/m³ greater than the surface
value. The mixed layer shoals by more than 20 m in the central gulf in 1977-1988 relative to 1970-1976, a reduction of ~30%. Unlike in temperate waters where there is a robust inverse relationship between SST and MLD anomalies [e.g. see Deser et al., 1996], the shoaling in the central GOA occurs near the nodal line for decadal SST changes (not shown), and ∆SST and ∆MLD are both positive along the North American coast.

The MLD in Feb and FMA are overlain on the temperature and salinity profiles in FMA from 1960 to 1999 averaged over a region (46°N-52°N, 160°W-140°W, box in Fig. 2) in the central GOA (Fig. 3). The MLD, which extends to the upper portion of the halocline, closely tracks the vertical variations in salinity but not temperature, including the epoch changes in the halocline around 1976. The close correspondence between MLD and the halocline occurs even though the salinity variations near the surface are negligible in the model due to the strong restoring of the surface salinity towards climatology. Furthermore, the epoch difference in salinity, at depths of 70 m to 150 m, closely resemble the MLD pattern, with increased salinity in the center of the gyre and decreased salinity along the coast (not shown). Our results, in conjunction with those of Cumnins and Lagerloef [2002] and Capotondi et al. [2005], suggest that the wintertime MLD in the GOA is primarily set by dynamical ocean processes that regulate the depth of the halocline. The shoaling of the halocline and hence the mixed layer in the central GOA, reflects the enhanced Ekman pumping associated with a deeper Aleutian low during 1977-88 relative to 1970-76 [see Fig. 4 in Capotondi et al., 2005].

Changes in MLD can influence primary productivity by regulating the amount of light and nutrients available for photosynthesis. In the northeast Pacific, the highest
primary productivity (PP) during the seasonal cycle, integrated over the upper 100 m of the model, occurs in March-May (see section 3.3). The epoch difference in primary productivity ($\Delta$PP in mmol N m$^{-2}$ day$^{-1}$) during MAM is also shown in Fig. 2. The $\Delta$PP pattern, with a decrease in the central GOA ringed by an increase along the coast, is remarkably similar to the structure of $\Delta$MLD; indeed the pattern correlation between the two is 0.91.

The positive correlation between MLD and PP initially suggested to us that vertical mixing of nutrients into the surface layer may have been the cause of the simulated biological changes in 1977-88 relative to 1970-76 in the GOA. However, the changes in nutrients do not appear to be responsible for the reduction in productivity after 1976, since the patterns of the $\Delta$NO$_3$ and $\Delta$H$_4$SiO$_4$ are quite different than $\Delta$PP (not shown), and the concentrations of both NO$_3$ and H$_4$SiO$_4$ exceed 5 mmol m$^{-3}$ over most of the northeast Pacific during 1977-1988 (not shown), which is sufficient to maintain rapid phytoplankton growth [Chai et al., 2002, 2003].

3.3 MLD, Light Regulation and Trophic Interactions

What is the primary factor linking the epoch difference in MLD with primary productivity in the GOA if it is not a change in (macro) nutrients? One possibility is that changes in MLD impact phytoplankton by altering the light available for photosynthesis, followed by trophic interactions that modify the initial biological response as the seasonal cycle progresses. Specifically, a shoaling of the wintertime MLD in 1977-88 relative to 1970-76 leads to more light and thus enhanced primary productivity and greater phytoplankton biomass earlier in the year in the central GOA region. The resulting
increase in phytoplankton enhances the food supply enabling a more rapid rise in the zooplankton biomass, but the associated increase in grazing subsequently suppresses phytoplankton and then zooplankton biomass during their peak in spring.

To test this hypothesis we first examine if the epoch changes are coherent across trophic levels and if these changes vary over the seasonal cycle. The 1977-88 mean and Δ in biomass are shown in Fig. 4a-d for each of the four plankton classes over the northeast Pacific. (The mean values during 1977-88 are similar to those from the entire 1960-1999 record - not shown). The biomass values (mmol N m⁻³) are presented during the calendar month in which the mean and Δ peak: March for small phytoplankton (P1), April for microzooplankton (Z1) and large phytoplankton (P2), and May for mesozooplankton (Z2). The biomass shown in Fig. 4 (and from hereon) is obtained from the surface layer (0-10 m) of the model, where the epoch differences are largest. The mean P1 and P2 biomass (contours) peaks at ~50°N, while the Z1 and Z2 reach maximum values north of ~52°N.

We note that the simulated long-term mean plankton biomass is greater than observed during the warm season, i.e. P is approximately twice that measured at Ocean Weather Station P (50°N, 145°W) during April-June [Chai et al., 2003]. Additionally, spring blooms are not commonly observed in the central GOA [Boyd and Harrison, 1999; Brickley and Thomas, 2004]. The overabundance of plankton, and P2 in particular, is likely due to the treatment of iron-limited growth in the model, as discussed further in section 4.

The Δ biomass for all four plankton classes (shading in Fig. 4) is negative and of large amplitude relative to the mean, indicating a substantial decrease in biomass after
1976. For example, P2 decreases by more than 60% in the vicinity of 46°N, 150°W during April. The reduction in both phytoplankton and zooplankton biomass are collocated and centered within and slightly to the south of the maximum in the long-term mean. The average biomass in the two periods and the difference between them are presented as a function of calendar month for the central GOA region in Fig. 4e and 4f, respectively. The P and Z biomass in the central GOA region decreases by ~40% in 1977-1988 relative to 1970-76 during its spring peak. This decrease is both preceded and followed by an increase in biomass in all four plankton classes, although the increase is relatively small except for P2 and Z2 in March and April, respectively. The spatial coherence of $\Delta P$ and $\Delta Z$, and the reversal in sign of the $\Delta$ in biomass with month is consistent with our hypothesis that trophic interactions and their evolution over the seasonal cycle play an important role in modulating the low-frequency ecosystem variability.

Given that the magnitude of the mean and the epoch difference in biomass are much greater for the larger plankton classes (Fig. 4), we focus on the dynamics of P2 and Z2 from hereon. The time series of the P2 and Z2 biomass (mmol N m$^{-3}$) from 1960-1999 averaged over the central Gulf of Alaska region for the months of March, April and May, along with the MLD (m) in March, are shown in Fig. 5. The MLD is well above average during the early epoch, except for 1970 and well below average for later epoch, except for 1977, 1980 and 1982, which are near normal. There is an inverse relationship between MLD and P2 biomass (Fig. 6a) as reflected by a -0.67 correlation between the two over the entire record (significant at the 99% level allowing for autocorrelation in the timeseries, e.g. see Wilks, 1995). The phytoplankton biomass in March is below normal
for all years between 1971 and 1976 but it is near normal and has much greater
interannual variability during the 1977-88 period, which suggests that the deeper mixed
layer during the first period has a greater impact on photosynthesis and phytoplankton
biomass than the shallower mixed layer in the later period. The asymmetry in the
relationship between MLD and P2 biomass is also indicated by a scatter plot of the two
variables averaged over the GOA region during March, where P is nearly independent of
MLD for depths < ~65 m (Fig 6a). The existence of a threshold depth, however, is
difficult to determine given the scatter in Fig. 6a, i.e. the P2 variability that is unrelated to
MLD, and the small number of relatively shallow MLD values.

In winter the phytoplankton and zooplankton biomass are low and vary together in
the GOA region (Fig. 5a), e.g. the correlation between P2 and Z2 in March is 0.72. The
positive correlation suggests that grazing is not controlling phytoplankton biomass
accumulation. While the Z2 biomass is generally low during March, it is nearly zero
during 1970-76 (Figs. 4e and 5a). Very low zooplankton abundance in winter can be a
critical factor in enabling rapid phytoplankton growth in spring [Fig 6c; Evans and
Parslow, 1985; Frost, 1991; Fasham, 1995]. Indeed, the P2 biomass switches from well
below normal in March to well above normal in April during 1970-76 (Fig. 5b). The Z2
biomass remains anomalously low in April, reflecting the limited food supply in the
previous month. Following the increase in P2 biomass in April, the Z2 biomass in May is
above normal during the early 1970s. In contrast to 1970-76, neither P2 nor Z2 exhibit
significant changes in the mean during 1977-88; rather, they exhibit large interannual
variability.
The lagged P-Z relationships during 1970-1976 in March through May are also very robust over the entire 40-year record, as indicated by the limited scatter between P2 and Z2 (Fig. 6b,c). The P2(Mar)-Z2(Apr) and P2(Apr)-Z2(May) correlations are 0.89 and 0.95, respectively, while the Z2(Mar)-P2(Apr) and Z2(Apr)-Z2(May) correlations are both ~-0.85. The relationship between P2 with Z2 in the following month appears to be linear (Fig. 6b), with a greater change in Z2(Apr) relative to a unit change in P2(Mar) than for Z2(May) relative to P2(Apr). While a linear relationship provides a reasonable fit for Z2(Mar)-P2(Apr) and for Z2(Apr)-P2(May), the two combined suggest a decreasing exponential function (Fig. 6c). The concurrent P2-Z2 correlations in the GOA region are -0.72 and 0.77 in April and May respectively; while significant, the change of sign between months and the slight decrease in magnitude compared to the lag correlations, suggest that the concurrent P-Z correlation values during spring reflect the lag relationship between the two trophic levels. However, we may not be fully resolving the period of the lag with monthly data.

The magnitude of the lead-lag correlations between P and Z at individual grid points generally exceeds 0.4 over the most of the northeast Pacific and 0.6 in the central and western GOA (not shown). Consistent with the regional analyses, the one-month lag P2-Z2 correlations are positive and the Z2-P2 correlations are negative for both March to April and April to May.

The timing of the mixed layer depth and plankton changes with respect to the seasonal cycle are examined further in Fig. 7 using time-latitude (Hovmöller) diagrams of monthly mean (contours) and Δ (shaded) values of a) MLD (m) and b) P2 and c) Z2 biomass (mmol N m⁻³). The variables are averaged over 160°W-140°W and presented for
40°N-57°N from January through July. The maximum mean MLD exceeds 80 m during January-March in the vicinity of 50°N and then decreases at all latitudes to about 20 m by May. The mean P2 biomass begins increasing in mid-winter, reaching a maximum in February, March, and April, for the latitude bands 40°-44°N, 44°-47°N and 48°-57°N, respectively. In addition to the seasonal cycle in insolation, the canonical evolution of P2 also depends on MLD, which is shallower in the southern part of the domain in JFM and thus more conducive for photosynthesis in a light-limited regime. The mean Z2 biomass increases during late winter, peaks in April-May and then declines but much more slowly than P2.

During January-March the \( \Delta \) P2 biomass increases where the \( \Delta \) MLD has decreased and vice versa, with the greatest increase in P2, located at 48°N in March, occurring in conjunction with the greatest decrease in MLD (shading, Fig. 7). This inverse relationship between \( \Delta \)MLD and \( \Delta \)P2, occurs over most of the domain during winter and is consistent with changes in light limitation influencing phytoplankton growth on decadal time scales. However, this relationship breaks down later in the spring and south of ~46°N in late winter as light is less of a factor in regulating primary productivity and grazing by zooplankton may have already begun to constrain the phytoplankton biomass.

As in the central GOA region, \( \Delta \)P2 and \( \Delta \)Z2 are enhanced prior to the annual mean peak in winter, reduced during and slightly after the peak in spring, and then weakly enhanced by July between 40°N and 54°N. For example, at 48°N the sharp increase in P2 in March is followed by a rapid rise in Z2 in April, while the decrease in P2 in April is followed by a decrease of Z2 in May.
Primary productivity (PP) of P2 and grazing (G) of P2 by Z2 (mmol N m\(^{-2}\) day\(^{-1}\)) averaged over the upper 100 m are shown for northeast Pacific during the months of February through May in Fig. 8. The maximum PP averaged over the 1977-88 period (contours) migrates from the southern to the northern edge of the GOA from February to May, as the amount of light necessary for photosynthesis moves northwards. This migration is not zonally uniform, as the largest values occur in the latitude bands 135\(^\circ\)W-140\(^\circ\)W, 130\(^\circ\)W-135\(^\circ\)W and 135\(^\circ\)W-150\(^\circ\)W in Feb, Mar and Apr-May, respectively. The overall maximum mean PP occurs in April rather than June (not shown), suggesting that factors other than the availability of light is limiting phytoplankton growth in late spring and summer. In general, the mean PP and G patterns are very similar, although the latter is shifted south of the former but only by ~2° latitude. The collocation of PP and G is consistent with local grazing on P2 by Z2, since the plankton life cycles occur much faster than the advection by ocean currents. The slight southward displacement of G relative to PP is likely due to the initiation of photosynthesis earlier in the seasonal cycle and thus further north relative to grazing by zooplankton and so the spatial pattern of G in March and April resemble the PP pattern in the previous month.

Like the mean values, the \(\Delta\)PP and \(\Delta\)G (shading in Fig. 8) also move northward across the GOA from February to May and the grazing is shifted slightly south relative to the primary productivity. \(\Delta\)PP and \(\Delta\)G are generally positive to the north and negative to the south of their respective maximum in the 1977-88 mean values, indicating a northward displacement in biological activity after the 1976 regime shift. The epoch differences are also a substantial fraction of the mean, e.g. primary productivity decreased by as much as ~70% in 1977-1988 relative to 1970-1976. There is generally a
close correspondence between the \( \Delta PP \) and \( \Delta G \) for each calendar month. Thus, there is both enhanced PP and G over much of the Alaskan gyre during February through April after 1976. The increase in grazing is especially pronounced between 48ºN-52ºN and 145ºW-155ºW in March and April, and likely contributes to the strong suppression of PP in May in the central GOA.

4. Summary and Conclusions

We have used a physical-ecosystem model to examine whether changes in the physical environment associated with the 1976-77 transition influenced the lower trophic levels of the food web and if so by what means. We hypothesize the following chain of events lead to the difference in the physical climate and biology in the central Gulf of Alaska in 1977-88 relative to 1970-76 in the model. The Aleutian Low strengthened [e.g. Trenberth and Hurrell, 1994] and the associated cyclonic winds accelerated the Alaskan gyre and enhanced Ekman pumping [Capotondi et al., 2005]. The resulting increase in upwelling caused the halocline to shoal in the center of the Alaskan Gyre. Thus the mixed layer, which extends to the upper portion of the halocline in winter, did not penetrate as deep. As a result, more phytoplankton remained in the euphotic zone and primary productivity and phytoplankton biomass increased earlier in spring. The enhanced food supply led to an increase in zooplankton biomass but then grazing pressure led to a strong (~40%) decrease in phytoplankton by April followed by a reduction in zooplankton by May. The one-month lag between P and Z in spring likely reflects the timescales set by photosynthesis and grazing, although this was based on monthly averaged data and does not consider all of the potential biological interactions in the system such as nutrient
recycling (see Fig. 1). Finally, there was a modest rebound in plankton concentrations by mid summer, but the magnitude of the simulated increase (~10%) over the GOA was substantially smaller than the observed doubling in zooplankton biomass [Brodeur and Ware, 1992].

The results presented here are dependent on the simulation of variability in both the physical and biological models. As in previous studies the winter mixed layer depth appears to be a critical variable for ecosystem dynamics in the North Pacific. Decadal variability of MLD in the northeast Pacific depends on dynamical ocean processes that influence the density jump at the base of the mixed layer (and top of the halocline). Interannual and decadal changes in MLD and phytoplankton biomass in the GOA are inversely related, indicative of a light regulated ecosystem as suggested by [Boyd et al., 1995, Polovina et al., 1995], but only during winter. Subsequent changes in grazing rates appear to cause larger changes in biomass during their spring peak, leading to a positive correlation between winter MLD and spring primary productivity/plankton biomass. The importance of grazing in regulating the spring plankton biomass in the northeast Pacific has been noted before in the context of the mean seasonal cycle using one-dimensional models [Evans and Parslow, 1985; Frost 1991; Fasham, 1995] and a three-dimensional physical/biogeochemical model [Gregg, 2002]. The epoch differences in biomass can also be viewed as changes in the seasonal cycle with the spring transition, or alternatively the northward seasonal advance of primary and secondary production, beginning earlier in the year in 1977-88 relative to 1970-76. Mackas et al. [1988], Stabeno and Overland [2001] and Bograd et al. [2002] have also found that decadal variability in the North Pacific climate and ecosystems can be manifest in the seasonal cycle.
Based on previous analyses [e.g. Bond et al., 2003] and the simulation of SST in the OGCM, we selected 1970-76 and 1977-88 as periods that exhibited “regime-like” behavior, i.e. when anomalies are of one sign. This appeared to be a fairly reasonable assumption for MLD depth, which is shallow (deep) during 1977-88 (1970-76) in the central GOA. However, the biological anomalies were consistent in sign only for 1970-76; during 1977-88 phytoplankton and zooplankton biomass were close to their long-term means and exhibited substantial interannual variability. The reason for this difference between periods is unclear, although it may be due to a nonlinear relationship between MLD and primary productivity, i.e. the phytoplankton is much more sensitive to changes in the March MLD, when the latter exceeds ~65 m in the GOA region. Perhaps when the MLD is less than a given depth enough light is available to sustain phytoplankton growth and it is thus not a strong controlling factor. The high degree of interannual variability during the later period also has important implications for assessing regimes based on observations: insufficient sampling could lead to inaccurate estimates of the actual mean value over a given period.

The model used in the present study may not adequately represent several potentially important processes, including eddies, near-shore processes such as coastally trapped waves, and zooplankton behavior and their interactions with higher trophic levels. In addition, the eastern subarctic Pacific is a high-nutrient, low-chlorophyll (HNLC) region where micronutrients, especially iron, are believed to limit the growth of phytoplankton in late spring and summer [Martin and Fitzwater, 1988; Boyd et al., 1996; SERIES, 2006]. The model’s implicit treatment of iron limitation may be adequate for the equatorial Pacific [Chai et al., 2002, 2007], but appears too weak in the northeast Pacific.
This may have lead to an overestimate of the simulated change in primary productivity and plankton biomass in response to the physical changes in the system. However, the processes identified here are still likely to play an important role in decadal variability in the GOA, since the lack of sunlight in conjunction with low iron levels may co-limit growth in winter [Maldanado et al., 1999], and grazing in combination with the limited iron supply may modulate spring blooms [Frost, 1991; Fasham, 1995; Aumont et al., 2003]. While the direct simulation of the iron cycle has recently been included in OGCMs [Aumont et al., 2003; Gregg et al., 2003; Moore et al., 2004], several key processes that influence iron cycling in the marine environment are poorly known and thus crudely parameterized. Questions remain regarding the bioavailability, cellular quotas and abiotic scavenging of iron and the magnitude of its external sources [Johnson et. al. 1997; Fung et al., 2002; Johnson et al., 2002]. Thus, the role of iron-limitation and small scale/coastal processes upon decadal variability in the North Pacific Ocean warrants further exploration.

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Figure Captions

Fig. 1. The epoch difference, 1977-88 – 1970-76 (Δ), in SST (°C) during January, February, March (JFM) from a) observations and b) the model simulation. The observed values are from COADS.

Fig. 2. The ΔMLD (m) (contours) and ΔPP (in N m$^{-2}$ day$^{-1}$) averaged over the upper 150 m of the ocean during JFM. The PP values can be converted from units of nitrogen to units of carbon by multiplying them by 6.625 as indicated by the Redfield ratio. Values have been spatially smoothed using a 9-point filter. The box indicates the central Gulf of Alaska (GOA) region.

Fig. 3. The a) temperature (°C) and b) salinity (ppt) over the upper 120 m of the ocean during FMA in the central GOA region (46°N-52°N, 160°W-140°W) for the years 1960-1999. The MLD during February and FMA are shown by an open square and closed circle, respectively.

Fig. 4. The 1977-88 mean (contours) and Δ (shading) for a) small phytoplankton (P1) in March, b) large phytoplankton (diatoms, P2) in April, c) microzooplankton (Z1) in April and d) mesozooplankton (Z2) in May. Also shown are the values of P1, P2, Z1, Z2 for each calendar month for the periods e) 1970-76 and 1977-88, and f) Δ, the difference between the two periods. The P and Z values (mmol N m$^{-3}$) presented here and the subsequent figures are from the top model level.

Fig. 5. The P2 and Z2 biomass (mmol N m$^{-3}$) from 1960-1999 averaged over the central Gulf of Alaska region for the months of a) March, b) April and c) May, along with the MLD (m, scale on right axis) in a) March. Also shown are the MLD, P2 and Z2 means (thin lines) over the 1960-1999 period.

Fig. 6. Scatter plot of a) MLD (March) with P2 (April), b) P2 (March) with Z2 (April) and P2 (April) with Z2 (May), c) Z2 (March) with P2 (April) and Z2 (April) with P2 (May). MLD (m) and P2 and Z2 (mmol N m$^{-3}$) values are for the central GOA region for the years 1960-99. Correlation (r) values between the variables are also given.

Fig. 7. Hovmöller (latitude-time) diagrams of the 1977-88 mean (contours) and Δ (shaded) values of a) MLD (m), b) P2 (mmol N m$^{-3}$) and c) Z2 (mmol N m$^{-3}$). The values are averaged over 160°W-140°W and shown for January through July.

Fig. 8. The 1977-88 mean (contours) and Δ (shaded) P2 primary productivity (left column) and grazing of P2 by Z2 (right column) for the months of February, March, April, and May. The values have been averaged over the upper 150 m and are in units of mmol N m$^{-2}$ day$^{-1}$. 
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MLD, Temperature, Salt in GOA Region
(46°N-52°N, 160°W-140°W) in FMA

Temperature (°C) and MLD

Salinity (ppt) and MLD

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