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Decadal variations in the California Current upwelling cells

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[1] We investigate decadal variations in the three-dimensional structure of the California Current System (CCS) upwelling cells as a potential mechanism for explaining observed ecosystem changes after the mid-1970s. To this end, we track the origin of upwelled water masses using adjoint passive tracers during time periods corresponding to the positive and negative phase of the Pacific Decadal Oscillation (PDO) in a 55 year regional ocean model simulation of the CCS. Results show that in the PDO "cool" phase (pre mid-1970s), the upwelling cell is deeper while during the "warm" phase (post mid-1970s), the upwelling cell is shallower with more horizontal entrainment of surface waters from the north. These changes in the coastal upwelling cell exhibit a latitudinal non-uniformity and may result in significant changes of the nutrient flux, which would have important impacts on the biological productivity of the coastal ocean. **Citation:** Chhak, K., and E. Di Lorenzo (2007), Decadal variations in the California Current upwelling cells, *Geophys. Res. Lett.*, *34*, L14604, doi:10.1029/2007GL030203.

1. Introduction

[2] A sharp decline in zooplankton biomass was observed off the coast of California after the mid-1970s [*Roemmich and McGowan*, 1995] and may have been related to the decline in the populations of many other marine animals [*Hare and Mantua*, 2000; *Chavez et al.*, 2003; *McGowan et al.*, 2003]. These ecosystem changes have been linked to large-scale climate variations in the North Pacific such as the Pacific Decadal Oscillation (PDO) [*Mantua et al.*, 1997; *Chavez et al.*, 2003], El Niño [*Bograd and Lynn*, 2001, 2003] and greenhouse warming [*Roemmich and McGowan*, 1995; *McGowan et al.*, 2003]. The decline in zooplankton biomass and other marine populations after the mid-1970s may be the result of changes in coastal upwelling caused by these large-scale climate variations since coastal upwelling impacts vertical nutrient fluxes along coastlines and is the most prominent mechanism thought to influence the populations of coastal ecosystems. In fact, the dramatic changes in the coastal ecosystems are coincident with warmer sea surface temperatures (SSTs) in the California Current System (CCS) associated with the 1976–1977 large-scale climate shift over the Pacific [*Miller et al.*, 1994] which may have affected ocean stratification such that coastal upwelling dynamics were influenced [*Di Lorenzo et al.*, 2005].

[3] Though the decline in zooplankton biomass and other marine populations after the mid-1970s may also be attributed to anomalous changes in the nutrient content of waters brought to the euphotic zone caused by a deeper thermocline [*Roemmich and McGowan*, 1995], this study focuses purely on changes in coastal upwelling (i.e., vertical velocities) as a *potential mechanism* for the aforementioned ecosystem changes. The goal of this study is to understand how climate variations in the North Pacific, specifically the PDO, influence the upwelling cell and the vertical fluxes of important tracers along the coast in the northeastern Pacific and to explore the potential impacts on coastal ecosystems. In this study, we use model adjoint passive tracers to elucidate how different phases of the PDO alter the three-dimensional upwelling cells in the CCS.

2. Model and Data

[4] The ocean model experiments are conducted with the Regional Ocean Modeling System (ROMS), a free-surface, hydrostatic, primitive equation ocean model with terrain-following coordinates [*Haidvogel et al.*, 2000; *Shchepetkin and McWilliams*, 2004]. ROMS has been widely used for regional and basin wide studies of the North Pacific Ocean [*Marchesiello et al.*, 2003; *Di Lorenzo*, 2003; *Capet et al.*, 2004; *Curchitser et al.*, 2005; *Di Lorenzo et al.*, 2005; *Moore et al.*, 2006; *Seo et al.*, 2007]. The adjoint ROMS (ADROMS) [*Moore et al.*, 2004] was also used to conduct an ensemble of passive tracer simulations.

[5] The ROMS computational domain covers (145°W–113°W, 27°N–51°N) (Figure 1) with an average resolution of 20 km in the horizontal and 30 levels in the vertical. To ensure that this resolution is adequate to capture the changes in the upwelling dynamics, selected portions of the simulations were repeated at 9 km resolution. These higher resolution simulations confirm the main features and results of the 20 km integrations, which are described in the next section.

[6] Open boundary and initial conditions are obtained from a basin-scale Northeast Pacific solution [*Combes and Di Lorenzo*, 2007]. Monthly mean (1950–2004) wind stress from the National Centers for Environmental Prediction are used as surface forcing. Monthly climatologies for heat and freshwater fluxes are computed from an 80-year spin-up integration that uses nudging to climatological sea surface temperature and salinity. For heat fluxes we add a nudging factor that relaxes the model SST towards time dependent (1950–2000) monthly mean SSTs from the National Oceanic and Atmospheric Administration [*Smith and Reynolds*, 2004].

3. Results

[7] In the 55-year ROMS simulation (1950–2004) of the CCS we select time periods corresponding to the negative (cold) and positive (warm) phase of the PDO (Figure 1e). Within these time periods, high SST anomalies events are used to generate the negative and positive phase ensembles.

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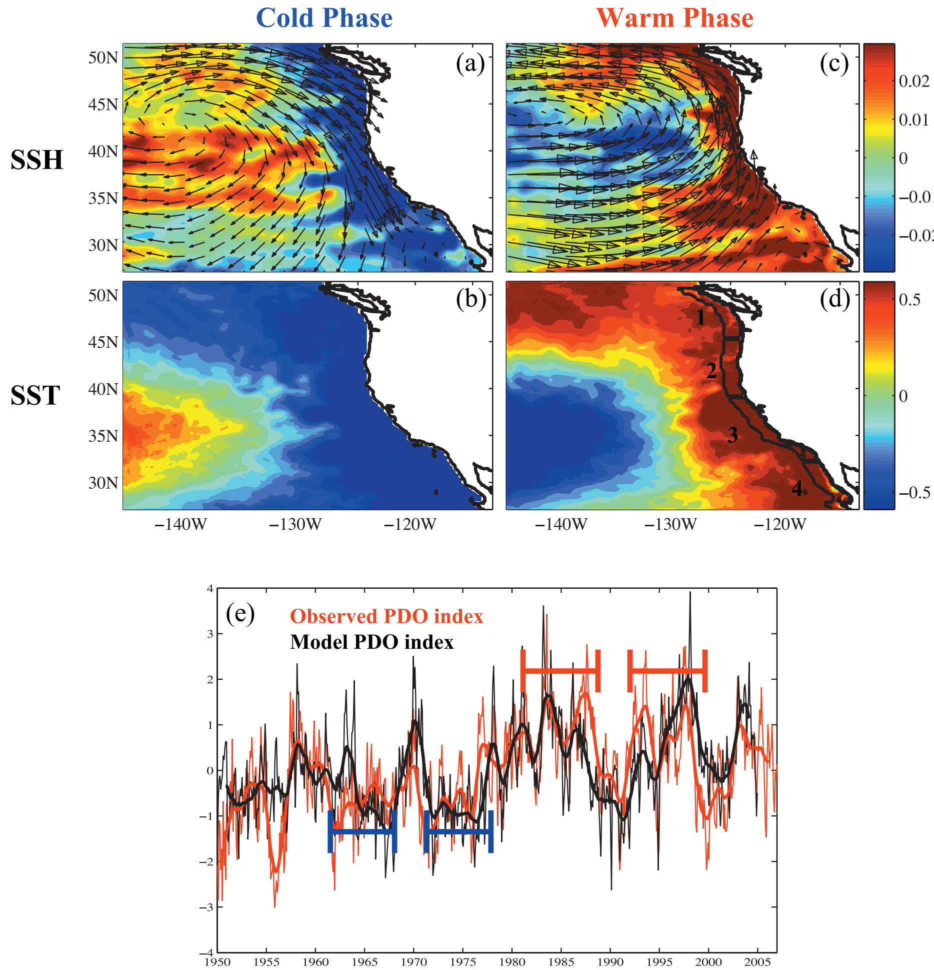

Figure 1. Ensemble mean anomalies of (a and c) SSH and (b and d) SST associated with the cold and warm phase of the PDO. Black arrows are the wind stress anomalies. The regions (1–4) of coastal upwelling used in this study are outlined in Figure 1d. (e) Time series of the "model PDO index" (black line) compared to the observed PDO index (red line). The horizontal blue and red lines in Figure 1e indicate the time periods from which years with high SST anomalies were extracted to comprise the cold and warm phase ensembles respectively.

The ensemble mean of model sea surface height (SSH), wind stress and SST anomalies for the negative phase (Figures 1a and 1b) show lower SSHs along the entire coastline and higher SSHs offshore, which is consistent with anomalous alongshore equatorward winds. The anomalous anti-cyclonic surface wind stress centered approximately near 140°W and 43°N (Figure 1a) also agrees with observations of a weaker Aleutian Low during a negative PDO [*Bond et al.*, 2003]. In the positive phase, we find higher SSHs (Figure 1c) and warmer SSTs (Figure 1d) along the coast and a tendency for more cyclonic surface wind stress. This is consistent with observations that show an intensification of the Aleutian Low [*Graham*, 1994] during a positive PDO, thus favoring more northward winds along the Pacific coast.

[8] The SSTa associated with the cold and warm phase ensembles (Figure 1) correlate strongly (R = ±0.87) with the first SST anomaly Empirical Orthogonal Function (EOF) of the model (not shown) and the Principal Component (PC) corresponding to this EOF correlates well (R = 0.63) with the observed PDO index. The PDO variability for the basin-scale model, which is used to prescribe the open boundary conditions, shows that SSH PC1 (the "model PDO index", Figure 1e), closely tracks the observed PDO (R = 0.82) [*Di Lorenzo et al.*, 2007].

[9] To determine how changes in the three-dimensional velocity field affect nutrient fluxes during the cold and warm phase, we use ensembles of passive tracer simulations from ADROMS. The evolution of passive tracers reflect advection and mixing and can be used to quantify the origin of water masses [*Vukicevic and Hess*, 2000; *Fukumori et al.*, 2004]. Each passive tracer simulation was carried out as follows. First, we identified the time of peak upwelling from monthly sea surface salinity anomalies in the CCS from the 55-yr simulation. At the onset of peak upwelling ($t = t_{up}$) in about mid-April, we tagged the coastally upwelled waters masses outlined in Figure 1d between ~0–30 m by injecting passive tracers for a duration of two weeks. Using ADROMS, we tracked the evolution of these passive tracers backwards in time to determine the origin of the coastally upwelled water masses one year prior to peak upwelling (t =

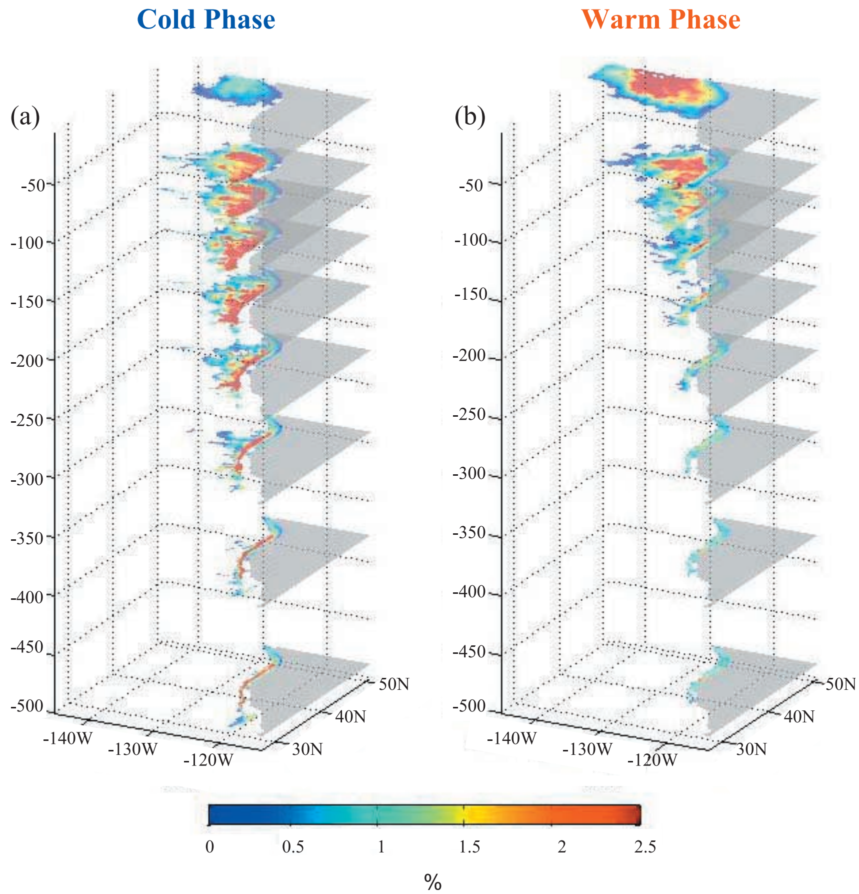

Figure 2. (a) Cold and (b) warm phase ensemble means of the percent ratio of passive tracer concentrations at time $t_{up\text{-}1}$ (one year prior to peak upwelling) relative to that at time t_{up} (the time of peak upwelling) indicating the origin of coastally upwelled waters (for region 2 in Figure 1d) one year prior to peak upwelling in mid-April.

$t_{up\text{-}1}$). Figure 2 shows the three-dimensional ensemble means of the percent ratio of passive tracer concentrations at time $t_{up\text{-}1}$ relative to that at time t_{up} associated with the cold (Figure 2a) and warm (Figure 2b) phase and effectively shows the origin of coastally upwelled waters (for region 2 in Figure 1d only) one year prior to peak upwelling. During the cold phase, much of the upwelled water mass originates from depths below 100 m indicating a deep upwelling cell, while during the warm phase, much of the upwelled water mass originates from above 100 m with most originating from the surface, indicating a shallow upwelling cell and a strong influence by lateral surface entrainment.

[10] The difference in the depth of upwelling cells between the cold and warm phase are not as striking for the southern regions (3 and 4 in Figure 1d) as they are for the northern regions (1 and 2 in Figure 1d) and is illustrated clearly in Figure 3 which shows two-dimensional representations of Figure 2 at varying depths for each coastal region (1–4 in Figure 1d). Compared to the northern regions, we see in the southern regions that there is less lateral surface entrainment and the passive tracer concentration varies little between the cold and warm phase below 100 m. This indicates that at lower latitudes, coastal upwelling cells in the CCS may be less affected by changes in wind stress associated with the PDO than at higher latitudes. Figure 3 also indicates that much of the upwelled waters come from offshore regions and from the north, especially at shallower depths (Figures 3b and 3f), where horizontal surface advection may still have an influence. At depths of around 200 m, the California Undercurrent (CU) influences water masses such that upwelled waters actually originate south of the region of upwelling (Figures 3c and 3d) for the different upwelling regions. The shoaling bathymetry near the coast may also cause upwelled waters to originate from offshore at these depths.

[11] The temporal evolution of the depth of coastal upwelling cells is similar to the temporal evolution of the model PDO index at low frequencies and is illustrated by Figure 4 which shows the normalized low-frequency time series of the anomalous passive tracer concentrations at $t_{up\text{-}1}$ at different depths of the ocean layer for region 1 and the model PDO index.

4. Discussion

[12] The adjoint passive tracer simulations used in this study are enlightening and provide a unique way of looking at variations in coastal upwelling dynamics. Specifically, the results of this study suggest a link between the PDO and the depth of coastal upwelling cells in the CCS. Our results indicate that nutrient rich deep waters are less likely to be vertically mixed to the surface during a positive or warm phase of the PDO compared to the negative or cold phase of the PDO. These nutrients are essential for the productivity

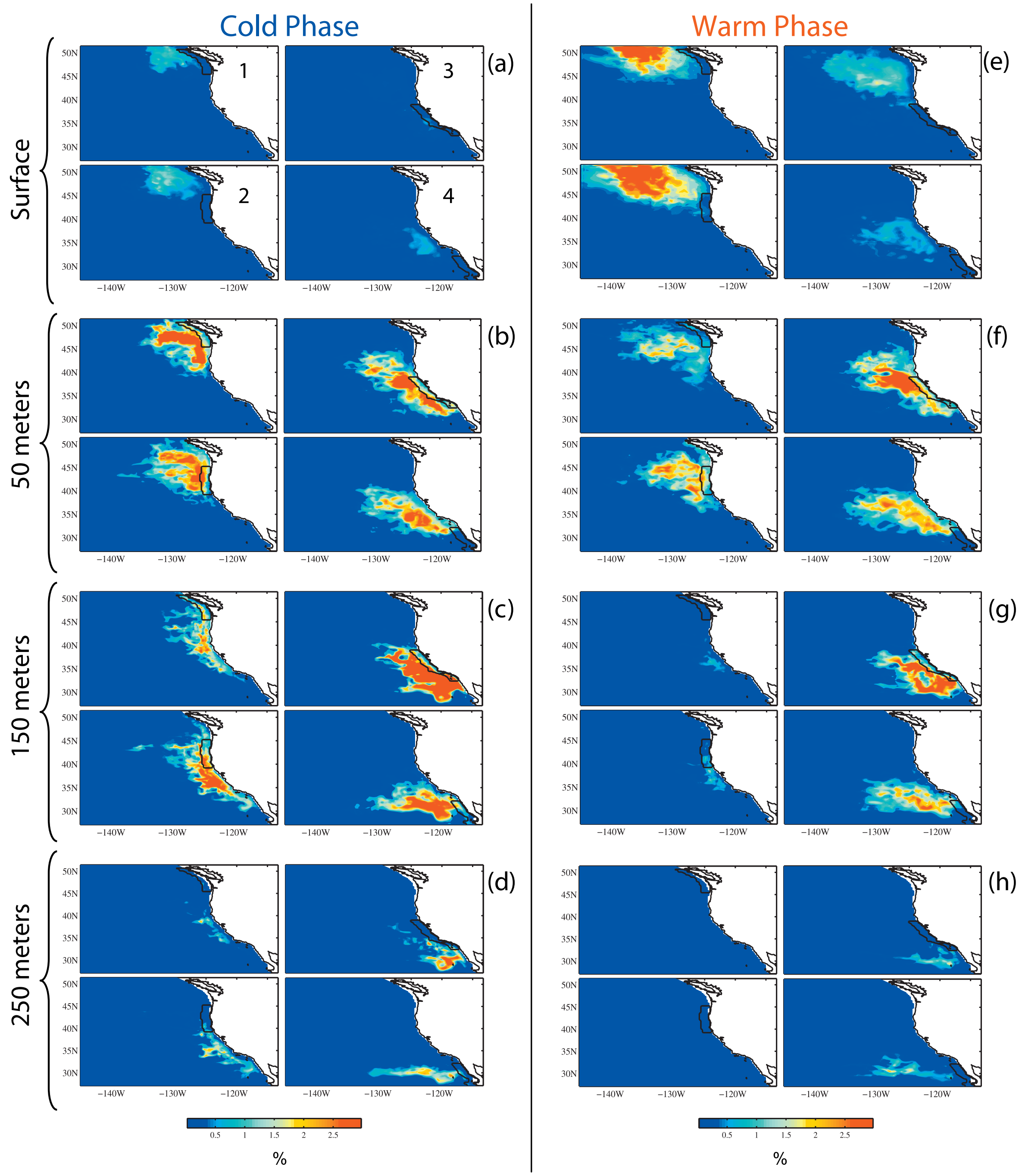

Figure 3. Same as Figure 2 for horizontal slices at (a and e) the surface, (b and f) 50 m, (c and g) 150 m, and (d and h) 250 m and all coastal regions (1–4 on Figure 1d). These slices indicate the origin of coastally upwelled waters one year prior to peak upwelling on two-dimensional planes.

of zooplankton and therefore, it is possible that the mid-1970's decrease in zooplankton biomass and other marine populations may *partly* be a reflection of the changes in upwelling dynamics associated with the climate shift. Observations that show dramatic changes in ecosystem productivity after the mid-1970s climate shift [*Ebbesmeyer et al.*, 1991; *Mantua et al.*, 1997; *Hare and Mantua*, 2000; *Brinton and Townsend*, 2003; *Chavez et al.*, 2003; *Lavaniegos and Ohman*, 2003] support this idea that nutrient fluxes important for productivity may be caused by decadal variations in the CCS upwelling cells. Additionally, observations of ecosystems [*King*, 2005] that show a non-uniform coastal response after the mid-1970s climate shift, support the idea that there may also be latitudinal variation in the CCS upwelling cells, which could be due to the latitudinal extent of the PDO "fingerprint".

[13] By repeating the experiment described above using only climatological winds or climatological SSTs (not shown), we find that it is not the ocean stratification, but rather the large-scale winds that are primarily responsible

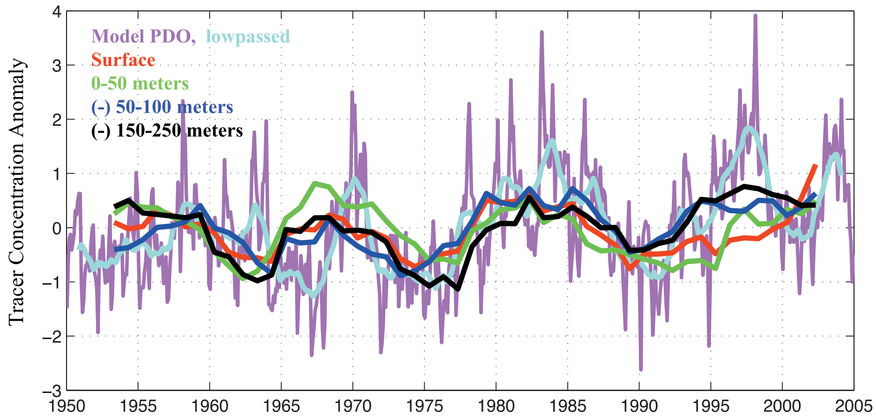

Figure 4. Time series of the low frequency normalized tracer concentration anomalies at the surface and between 0–50 meters, 50–100 meters (inverse is plotted), and 150–100 meters (inverse is plotted) at $t_{up\text{-}1}$ (one year prior to peak upwelling) indicating the originating depth of upwelled water masses. The time series of the model PDO index (raw in magenta, low frequency in cyan) are also plotted.

for the changes in coastal upwelling depth between the two different climate regimes. Therefore, the SST anomalies associated with the cold and warm phase ensembles are more likely just a reflection, rather than the cause, of the changes in upwelling dynamics. We speculate that changes in the ocean stratification caused by variations in the SST may, however, result in a positive feedback allowing the upwelling dynamics to perpetuate until the large-scale winds change again with respect to changes in the PDO. The following scenario is thus possible. During the cold phase, or negative phase of the PDO, the large scale climate is largely influenced by a weakened Aleutian Low which favors more northerly coastal wind stress resulting in cooler SSTs due to deeper upwelling and cooler, nutrient rich deep waters being vertically mixed to the surface. Cooler SSTs can lead to a less stratified ocean that encourages vertical mixing to continue (positive feedback) until the climate shifts to a warm phase (i.e., after the mid-1970s), where the opposite scenario occurs (strong Aleutian Low). The major source of cooler SSTs during the warm phase is from surface waters from the north that are advected southwards by the California Current, which has only been weakened by the intensified Aleutian Low.

[14] Though this study establishes a link between the PDO and CCS upwelling cell and decadal changes in mixing pathways, it also raises more issues and questions. The variation in the population of ecosystems is complex and depends not only on obtaining the needed nutrients but also many other physical variables (i.e., sunlight, amount of oxygen, etc.) and the ecosystem itself may also generate its own intrinsic variability independent of large-scale climate variability. Anomalous changes in the nutrient content of source waters, a topic not addressed here, may also have an impact on ecosystems [*Di Lorenzo et al.*, 2007]. Therefore, the variation in upwelling cell with the PDO only partly explains fluctuations in coastal ecosystems. Variations in the coastal upwelling cell are not only caused by the PDO. Global warming, which is projected to increase coastal upwelling [*Bakun*, 1990; *Snyder et al.*, 2003], may act in concert with or even influence the PDO, thereby affecting the variation in coastal dynamics explored in this study, though it is unclear how or even if global warming would influence the PDO. Additionally, our results indicate a non-uniformity in the coastal upwelling response to the PDO. Given the latitudinal location of the southern coastal upwelling cells, mesoscale eddies and perhaps El Niño may have more of an impact on the coastal dynamics in these southern regions. Work is currently underway to address some of these issues.

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References

Bakun, A. (1990), Global climate change and intensification of coastal ocean upwelling, *Science*, *247*, 198–201.

Bograd, S. J., and R. J. Lynn (2001), Physical-biological coupling in the California Current during the 1997–99 El Niño-La Niña cycle, *Geophy. Res. Lett.*, *28*, 275–278.

Bograd, S. J., and R. J. Lynn (2003), Long-term variability in the southern California Current System, *Deep Sea Res., Part II*, *50*, 2355–2370.

Bond, N. A., J. E. Overland, M. Spillane, and P. Stabeno (2003), Recent shifts in the state of the North Pacific, *Geophys. Res. Lett.*, *30*(23), 2183, doi:10.1029/2003GL018597.

Brinton, E., and A. E. Townsend (2003), A description of abundances of euphausiid species during 1950–2001 in the California Current System, *Deep Sea Res., Part II*, *50*, 2449–2472.

Capet, X. J., P. Marchesiello, and J. C. McWilliams (2004), Upwelling response to coastal wind profiles, *Geophys. Res. Lett.*, *31*, L13311, doi:10.1029/2004GL020123.

Chavez, F. P., J. Ryan, S. E. Lluch-Cota, and M. Ñiquen (2003), From anchovies to sardines and back: Multidecadal change in the Pacific Ocean, *Science*, *299*, 217–221.

Combes, V., and E. Di Lorenzo (2007), Asymmetric response to climate forcing of mesoscale eddies in the Gulf of Alaska gyre: Forced vs. intrinsic, *Prog. Oceanogr.*, submitted.

Curchitser, E. N., D. B. Haidvogel, A. J. Hermann, E. L. Dobbins, T. M. Powell, and A. Kaplan (2005), Multi-scale modeling of the North Pacific Ocean: Assessment and analysis of simulated basin-scale variability (1996–2003), *J. Geophys. Res.*, *110*, C11021, doi:10.1029/2005JC002902.

Di Lorenzo, E. (2003), Seasonal dynamics of the surface circulation in the Southern California Current System, *Deep Sea Res., Part II*, *50*, 2371–2388.

Di Lorenzo, E., A. J. Miller, N. Schneider, and J. C. McWilliams (2005), The warming of the California Current: Dynamics and ecosystem implications, *J. Phys.Oceanogr.*, *35*, 336–362.

Di Lorenzo, E., N. Schneider, P. J. Franks, S. Bograd, A. J. Miller, P. Riviere, and K. Chhak (2007), North Pacific gyre-scale oscillation controls salinity and nutrients in the California Current, *Science*, in prep.

Ebbesmeyer, C. C., D. R. Cayan, D. R. McLain, F. H. Nichols, D. H. Peterson, and K. T. Redmond (1991), 1976 step in Pacific climate: Forty environmental changes between 1968–1975 and 1977–1984, in *Proceedings of the Seventh Annual Pacific Climate (PACLIM) Workshop, Asilomar, California—April 1990: Climate Variability of the Eastern North Pacific and Western North America*, *Tech. Rep. 26*, edited by J. L. Betancourt and V. L. Tharp, pp. 120–141, Interagency Ecol. Stud. Program for the Sacramento, San Joaquin Estuary, Sacramento, Calif.

Fukumori, I., T. Lee, B. Cheng, and D. Menemenlis (2004), The origin, pathway, and destination of Niño-3 water estimated by a simulated passive tracer and its adjoint, *J. Phys. Oceangr.*, *34*, 582–604.

Graham, N. E. (1994), Decadal scale variability in the 1970's and 1980's: Observations and model results, *Clim. Dyn.*, *10*, 60–70.

Haidvogel, D. B., H. G. Arango, K. Hedstrom, A. Beckmann, P. Malanotte-Rizzoli, and A. F. Shchepetkin (2000), Model evaluation experiments in the North Atlantic basin: Simulations in nonlinear terrain-following coordinates, *Dyn. Atmos. Oceans*, *32*, 239–281.

Hare, S. R., and N. J. Mantua (2000), Empirical evidence for North Pacific regime shifts in 1977 and 1989, *Prog. Oceanogr.*, *47*, 103–145.

King, J. R. (Ed.) (2005), Report of the study group on fisheries and ecosystem response to recent regime shifts, *Sci. Rep. 28*, North Pac. Mar. Sci. Org., Sidney, B. C., Can. (Available at http://www.pices.int/)

Large, W. G., J. C. McWilliams, and S. C. Doney (1994), Oceanic vertical mixing: A review and a model with a nonlocal boundary-layer parameterization, *Rev. Geophys.*, *32*, 363–403.

Lavaniegos, B. E., and M. C. Ohman (2003), Long term changes in pelagic tunicates of the California Current, *Deep Sea Res., Part II*, *50*, 2493–2518.

Mantua, N. J., S. R. Hare, Y. Zhang, J. M. Wallace, and R. C. Francis (1997), A Pacific interdecadal climate oscillation with impacts on salmon production, *Bull. Am. Meteorol. Soc.*, *78*, 1069–1079.

Marchesiello, P., J. C. McWilliams, and A. F. Shchepetkin (2003), Equilibrium structure and dynamics of the California Current System, *J. Phys. Oceanogr.*, *33*, 753–783.

McGowan, J. A., S. J. Bograd, R. J. Lynn, and A. J. Miller (2003), The biological response to the 1977 regime shift in the California Current, *Deep Sea Res., Part II*, *50*, 2567–2582.

Miller, A. J., D. R. Cayan, T. P. Barnett, N. E. Graham, and J. M. Oberhuber (1994), The 1976–77 climate shift of the Pacific Ocean, *Oceanography*, *7*, 21–26.

Moore, A. M., H. G. Arango, A. J. Miller, B. D. Cornuelle, E. Di Lorenzo, and D. J. Neilson (2004), A comprehensive ocean prediction and analysis system based on the tangent linear and adjoint components of a regional ocean model, *Ocean Modell.*, *7*, 227–258.

Moore, A., H. Arango, E. Di Lorenzo, B. D. Cornuelle, and A. J. Miller (2006), An adjoint sensitivity analysis of the southern California Current circulation and ecosystem. Part I: The physical circulation, *J. Phys. Oceanogr.*, submitted.

Roemmich, D., and J. A. McGowan (1995), Climatic warming and the decline of zooplankton in the California Current, *Science*, *267*, 1324–1326.

Seo, H., A. J. Miller, and J. O. Roads (2007), The Scripps Coupled Ocean-Atmosphere Regional (SCOAR) model with applications in the eastern Pacific sector, *J. Clim.*, *20*(3), 381–402.

Shchepetkin, A. F., and J. C. McWilliams (2004), The Regional Oceanic Modeling System (ROMS): A split explicit, free-surface, topography-following-coordinate oceanic model, *Ocean Modell.*, *9*, 347–404.

Smith, T. M., and R. W. Reynolds (2004), Improved extended reconstruction of SST (1854–1997), *J. Clim.*, *17*(12), 2466–2477.

Snyder, M. A., L. C. Sloan, N. S. Diffenbaugh, and J. L. Bell (2003), Future climate change and upwelling in the California Current, *Geophys. Res. Lett.*, *30*(15), 1823, doi:10.1029/2003GL017647.

Vukicevic, T., and P. Hess (2000), Analysis of tropospheric transport in the Pacific Basin using the adjoint technique, *J. Geophys. Res.*, *105*, 7213–7230.

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