Decadal-Scale Climate and Ecosystem Interactions in the North Pacific Ocean

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Decadal-scale climate variations in the Pacific Ocean wield a strong influence on the oceanic ecosystem. Two dominant patterns of large-scale SST variability and one dominant pattern of large-scale thermocline variability can be explained as a forced oceanic response to large-scale changes in the Aleutian Low. The physical mechanisms that generate this decadal variability are still unclear, but stochastic atmospheric forcing of the ocean combined with atmospheric teleconnections from the tropics to the midlatitudes and some weak ocean-atmosphere feedbacks processes are the most plausible explanation. These observed physical variations organize the oceanic ecosystem response through large-scale basin-wide forcings that exert distinct local influences through many different processes. The regional ecosystem impacts of these local processes are discussed for the Tropical Pacific, the Central North Pacific, the Kuroshio-Oyashio Extension, the Bering Sea, the Gulf of Alaska, and the California Current System regions in the context of the observed decadal climate variability. The physical ocean-atmosphere system and the oceanic ecosystem interact through many different processes. These include physical forcing of the ecosystem by changes in solar fluxes, ocean temperature, horizontal current advection, vertical mixing and upwelling, freshwater fluxes, and sea ice. These also include oceanic ecosystem forcing of the climate by attenuation of solar energy by phytoplankton absorption and atmospheric aerosol production by phytoplankton DMS fluxes. A more complete understanding of the complicated feedback processes controlling decadal variability, ocean ecosystems, and biogeochemical cycling requires a concerted and organized long-term observational and modeling effort.

1. Introduction

As observations of the Pacific Ocean physical-biological system accumulate through extensive networks of observing systems, prominent fluctuations on decadal[†]

timescales increasingly reveal distinct linkages between the physical ocean climate and the oceanic ecosystem. It is imperative to develop a clear understanding of the mechanisms that lead to these physical-biological changes because of the important economical and societal impacts that these variations wield on our global communities.

Our objective here is to summarize key aspects of the current state of understanding of the processes controlling physical climate decadal variations of the Pacific Ocean, the oceanic ecosystem changes that have been observed to occur in the Pacific with similar timescales, and the mechanisms by which the physical environmental changes may alter the oceanic ecosystem and vice versa.

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[†] We consider climate variations with timescales longer than interannual (ENSO) and shorter than centennial (greenhouse-gas forcing) to have "decadal" timescales.

2. Observed Physical Climate Variability and Model Interpretations

2.1 Dominant large-scale patterns of decadal variability

Statistical analysis of the observed spatial and temporal variations of Pacific Ocean physical fields normally center around sea-surface temperature (SST) and sea-level pressure (SLP) because these have been routinely observed over the longest periods of time. The dominant atmospheric pattern that arises on decadal timescales is associated with changes in the Aleutian Low during wintertime (Trenberth and Hurrell, 1994; Parrish *et al.*, 2000). Winter is the season that atmospheric forcing of the ocean is strongest and that oceanic response fields persist the longest (Alexander *et al.*, 1999). A dominant pattern in SST variability (Fig. 1(a)) occurs in response to direct surface forcing (via Ekman currents, turbulent mixing, and surface heat fluxes) by the atmospheric wind stress fields associated with changes in the Aleutian Low (Cayan, 1992; Miller *et al.*, 1994a). This pattern has been identified in many previous studies (e.g., Zhang *et al.*, 1997; Enfield and Mestas-Nunez, 1999; Mochizuki and Kida, 2003) and it appears on seasonal, interannual, and decadal timescales (Tanimoto *et al.*, 1992). It features oppositely signed SST anomalies, viz., for a strengthened Aleutian Low SST exhibits a cold region in the central North Pacific surrounded to east, north and south by a warm pattern along the coast of western North America, northwards across the Gulf of Alaska and southwards across the eastern subtropics and tropics (Fig. 1(a)). Depending on the statistical analysis technique that is used, weaker and less consistent SST anomalies correlated with this pattern appear in various other parts of the Pacific.

This "canonical" pattern of SST response to atmospheric forcing projects strongly onto the first empirical orthogonal function (EOF) mode of SST over the region north of 20°N which has been dubbed the Pacific Decadal Oscillation (PDO, Mantua and Hare, 2002). But another pattern of SST response arises from a different aspect of the ocean's response to changes in the Aleutian Low (Deser and Blackmon, 1995; Nakamura *et al.*, 1997; Yasuda and Hanawa, 1997; Miller *et al.*, 1998; Xie *et al.*, 2000; Seager *et al.*, 2001; Tomita *et al.*, 2003). Changes in wind-stress curl (Ekman pumping driven by convergences and divergences in open-ocean Ekman currents), associated with the Aleutian Low variations, force the North Pacific gyre scale thermocline (Fig. 1(c)) and circulation to change (Miller *et al.*, 1998; Deser *et al.*, 1999; Yasuda and Kitamura, 2003). Part of this circulation adjustment process involves the propagation of baroclinic Rossby waves from the central North to the region around the subarctic frontal zone and the Kuroshio-Oyashio Extension (KOE) region. These Rossby waves arrive several years after the Aleutian Low changes, resulting in a lagged response in SST in the KOE region (Schneider and Miller, 2001). This second SST pattern (Fig. 1(b)) also projects onto the PDO although it has a lagged relationship to the canonical pattern of SST. Hence, the PDO index should be considered an amalgam of these two physical ocean responses.

2.2 Mechanisms of decadal variability

The mechanisms by which the Aleutian Low might vary on decadal timescales are uncertain, but stochastic excitation by midlatitude weather disturbances certainly exerts a strong influence. The stochastic forcing climate models of Hasselmann (1976) and Barsugli and Battisti (1998), for SST, and Frankignoul *et al.* (1997), for ocean currents, constitute zero-order models of midlatitude decadal variability. However, deviations (i.e., spectral peaks, or enhancements of variance) from the simple spectra predicted by those theories are significant, and many ideas have been advanced to explain them. Some of these involve spectral peaks arising with only stochastic atmospheric forcing (Jin, 1997; Saravanan and McWilliams, 1998; Neelin and Weng, 1999; Qiu, 2003), while others rely on feedback loops or remote forcing.

Deterministic forcing from the tropics clearly has an effect in establishing decadal SST variability in the midlatitudes. The forcing of the canonical SST pattern (around the subtropical front) has long been linked to atmospheric teleconnections from the tropics associated with El Niño/Southern Oscillation (ENSO) events on interannual timescales (Alexander *et al.*, 2002). The forcing of the KOE SST pattern (around the subarctic front) is only weakly linked to tropical teleconnections, because the nodal line of the ENSO-forced pattern runs through the KOE region (e.g., Hanawa *et al.*, 1988). The independent behavior of the KOE SST has been clarified through its enhanced decadal variance relative to the canonical SST pattern (Deser and Blackmon, 1995; Nakamura *et al.*, 1997) and through its links to decadal wind-stress curl forcing (Deser *et al.*, 1996, 1999; Miller *et al.*, 1998). Recent studies with a simple first-order Markov model with forcing specified by the tropical SST index, damping rate specified by SST persistence (with re-emergence) and white noise forcing (simulating midlatitude weather) reveals that the bulk of the PDO index is explicable by atmospheric forcing from tropical teleconnections (Newman *et al.*, 2003). The forcing with tropical origins (Graham *et al.*, 1994; Giese and Carton, 1999; Schwing *et al.*, 2002b; Yasunaka and Hanawa, 2003; Deser *et al.*, 2003b) clearly drives the canonical SST pattern portion of the PDO. However, the simple model result is somewhat deficient in decadal timescale energy. This suggests that the KOE SST pattern portion of the PDO is not simply driven by (or at least is not in phase with) this tropical forcing. Adding a lagged KOE response
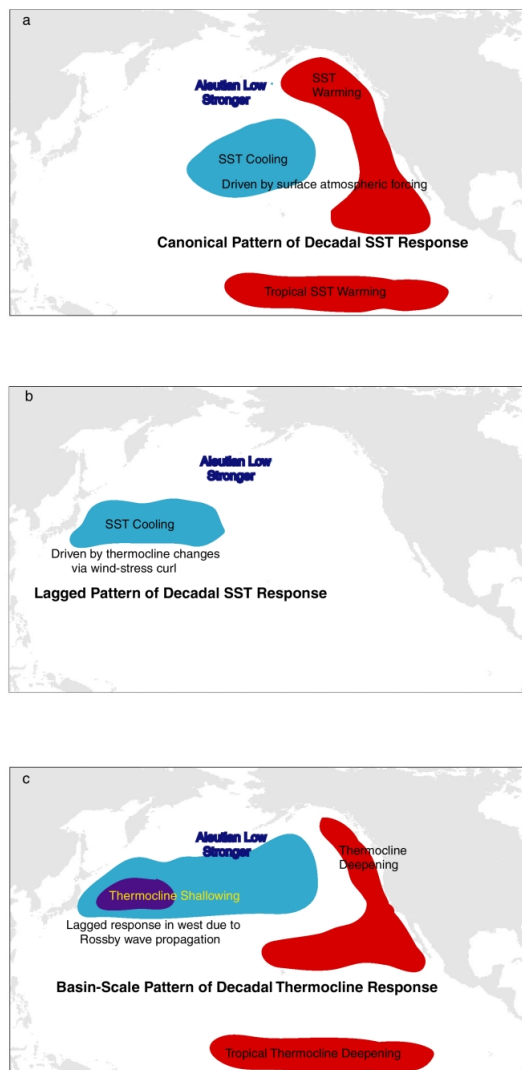
Fig. 1. (a) Schematic diagram of the canonical pattern of decadal SST response in the North Pacific. This pattern is largely controlled by, and is in phase with, surface forcing (Ekman current advection, vertical mixing, and surface heat fluxes) by the wind stress associated with a decadal change in the Aleutian Low. It often has a strong correlation with tropical Pacific SST due to atmospheric and oceanic teleconnections. (b) Schematic diagram of the KOE SST pattern of decadal SST response in the North Pacific. This pattern is largely controlled by the thermocline response to wind-stress curl forcing associated with a decadal change in the Aleutian Low. It has a several-year phase lag with the response in (a) due to Rossby wave propagation from the central North Pacific to the KOE region. (c) Schematic diagram of the basin-scale thermocline response pattern to wind-stress curl forcing associated with a decadal change in the Aleutian Low. It tends to be in phase with the (a) in the eastern basin and in phase with (b) in the western basin. Anomalous geostrophic currents follow the thermocline gradients around the gyres.
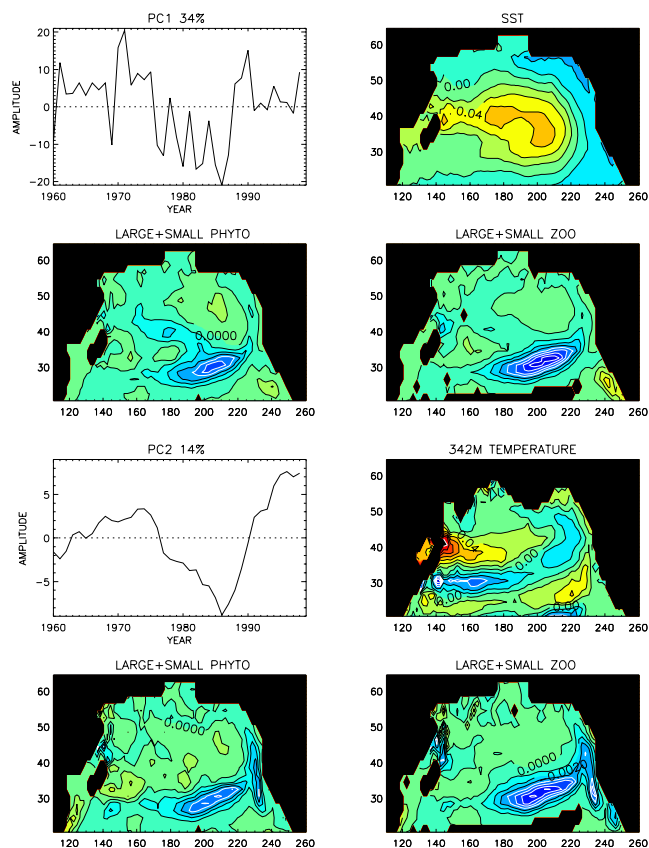

Fig. 2. (Top four panels) First combined EOF of anomalies of SST, small phytoplankton, large phytoplankton, small zooplankton, and large zooplankton from the Chai *et al.* (2003a) 1960–1999 hindcast driven by NCEP reanalysis winds. The biological variables are 0–100 m averaged, annual mean quantities (CI = 0.0005), and the SST (CI = 0.02) is wintertime (JFM) mean. Multiply contoured values by time coefficient to obtain dimensional values of °C and mmol nitrogen. After computing the EOF, the small and large biological variables were summed to reduce the plot complexity and because they were spatially similar. (Bottom four panels) As in top four panels, but for second combined EOF of annual mean 342 m temperature (CI = 0.02), a proxy for the thermocline depth, and biological variables (CI = 0.001). The first combined EOF with 342 m temperature explains a trend, not the decadal variability. N.b., the two combined EOF calculations for SST and 342 m temperature are computed independently.

pattern, mimicking the gyre-scale spin-up delay, may improve the fit of the Newman *et al.* (2003) simple model. Alternatively, midlatitude ocean-atmosphere, or ocean-atmosphere-ecosystem, feedbacks may be important.

Associated with atmospheric teleconnections are oceanic teleconnections via Kelvin wave-like disturbances emanating from the tropics and moving northwards along the North American coast (Meyers *et al.*, 1996; Clarke

and Lebedev, 1999). These mainly affect the coastal thermocline and seem to have little influence on large-scale midlatitude SST or the overlying atmosphere.

Sources for decadal variability in the tropical Pacific are then, of course, necessary to explain this determinis-tic forcing signal for the midlatitudes. Delayed oscillator models involving Rossby waves at higher latitudes (White et al., 2003) or having higher vertical mode numbers (Liu et al., 2003) have been shown to contain decadal varia-tions. Fundamentally non-linear theories can produce decadal timescales in simple models (Timmermann and Jin, 2002a).

Variations in spiciness* along tropical-subtropical subduction paths can occur in a full-physics coupled model in what appears to be coupled mode involving sub-tropical wind stress curl driving spiciness changes in the thermocline (Schneider, 2000). The spiciness changes in water upwelled along the equator can alter both the mean state of the tropical coupled ocean-atmosphere system as well as the magnitude and frequency of ENSO variations (Schneider, private communication, 2002). This type of tropical-subtropical interaction is a modified version of the Gu and Philander (1994) hypothesis, which was origi-nally proposed to link surface-forced SST of the midlatitude North or South Pacific to the tropical thermocline via equatorward subduction along isopycnal surfaces. While neither observations (Schneider et al., 1999) nor full-physics coupled models seem to contain oscillations explicable in that original oceanic subduc-tion theoretical framework, various other related pathways are still under investigation (e.g., Lysne et al., 1997; Liu, 1999; Kleeman et al., 1999; Liu and Yang, 2003; Solomon et al., 2003).

The midlatitude ocean may generate feedbacks to the overlying atmosphere resulting in increased variance at decadal timescales or possibly resulting in coupled decadal modes. Schneider et al. (2002) showed that the KOE SST field in the coupled model originally studied by Latif and Barnett (1994) drives a local atmospheric response through surface heat fluxes. The atmospheric response is evident in rainfall fields over the ocean but does not close a delayed oscillator feedback loop as Latif and Barnett (1994) originally suggested even though Rossby waves allow the KOE region to respond with a several-year lag to wind-stress curl changes. Instead, a positive feedback from ocean to atmosphere results in enhanced variance at decadal timescales. Even without invoking ocean feedbacks, Qiu (2003) showed that KOE zonal currents can oscillate with 10-yr periods due to the

latitudinal dependence of Rossby wave phase speeds. It is important to note that the atmosphere is very sensitive to changes in the KOE SST because the storm tracks pass over this region (Peng et al., 1997). The atmosphere is not very sensitive to the canonical SST pattern, which instead is essentially a forced, slave-like ocean response.

Variations of the Aleutian Low are also partly linked to the Arctic Oscillation (AO), also called the Northern Hemisphere Annular Mode (Thompson and Wallace, 2000). This atmospheric circulation pattern is a quasi-zonal flow that has variability on many climatic timescales and is much more strongly weighted in the North Atlan-tic than the North Pacific, and it was traditionally called the North Atlantic Oscillation (Hurrell, 1995). Although the signature of the AO in the North Pacific is rather weak, it is significant (e.g., Overland et al., 1999; Yasunaka and Hanawa, 2002). The source of this correlation is not known and may be due to AO forcing of the Aleutian Low, North Pacific sector forcing of the AO or a common re-mote forcing (e.g., Yasunaka and Hanawa, 2003).

The energy that forces the Earth's climate is derived from the sun, and ocean biology depends upon sunlight as well. The solar "constant" is not as its name implies but can vary significantly over time. Unfortunately, present solar flux observations are not available for the time scale required to access decadal variability. Satel-lite data sets are only now coming into long enough time lengths to be of use for short term studies (Scott and Alexander, 1999), and recent longer term (1950–1990) short wave budgets for the ocean have only resolved the effects due to decadal cloud variability (Moisan and Niiler, 1998).

Variability in the magnitude of the solar "constant" can occur through a number of sun related processes. Most notable of these is the 11-year sunspot cycle. However, additional modulations in the intensity of the sunspot cycle by the 90-year Gleissberg cycle (Eddy, 1977) may also play a role in climate modulations. The specific ef-fects that this has on the Earth's climate have been postu-lated by a number of researchers (Eddy, 1977) but have only recently been revealed through climate studies. More recently, Landscheidt (1999, 2001) has argued that the PDO is controlled by variations in the solar forcing. In a recent analysis of a 112 year coral-based sea surface tem-perature reconstruction (Cobb et al., 2001), a similar 12–13 year periodicity was observed. Of importance to cli-mate studies is the direct link between total cloud cover and cosmic ray flux that is known to support cloud for-mation (Svensmark and Friis-Christensen, 1997). Cloud dynamics remain one of the main uncertainties in climate modeling studies. A recent study by White et al. (1997) found that global average upper ocean temperatures re-sponded to changes in solar irradiance at periods of 9–13, 18–25, and >100 years. These frequencies closely

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*Spiciness is the coordinate orthogonal to density on a T-S diagram, such that high spiciness indicates hot and salty wa-ter relative to water of the same density that is cooler and fresher.

match those oscillations observed in varve thickness data analyzed by Biondi *et al.* (1997). In addition, the Maunder-Minimum, a known period of low solar flux, coincides with the period of transition in the Santa Barbara varve record indicating that solar variability may be an important modulator of the ecosystem variability at even greater time scales.

An additional possible source of climate modification is that the ultraviolet part of the solar spectrum undergoes larger variations than that for the longer wavebands (Friis-Christensen, 2001). In addition, spectral variations in cloud transmission may increase the fraction of the visible-ultraviolet radiance portion of the solar spectrum impacting the ocean surface, allowing the solar energy under cloudy skies to attenuate less rapidly than on clear sky days and increasing the impact of phytoplankton attenuation in modifying the solar heat flux into the ocean (Siegel *et al.*, 1999). While this may present itself as a secondary effect, in regions such as the Equatorial and North Pacific, where clouds and phytoplankton blooms are highly coupled to ENSO and decadal dynamics, respectively, it may play an important role as a feedback mechanism. Also, the link between phytoplankton production of dimethylsulfide (DMS) and clouds has been repeatedly argued as a strong feedback link for long-term climate variability (Charlson *et al.*, 1987), and may contribute to decadal variability as well (Miller *et al.*, 2003).

3. Forcing the Oceanic Ecosystem by Physical Climate

Physical forcing of decadal variations of the oceanic ecosystem has long been recognized as an important factor affecting phytoplankton, zooplankton, and higher order predators in the food web (e.g., Francis *et al.*, 1998). It remains unclear, however, whether the dominant part of the ecosystem response to climate forcing is quasi-linear or if the ecosystem responds parametrically to weak climate modulation through sensitive intrinsic nonlinearities in self-interactions of the biological components. The large-scale patterns of oceanic response to atmospheric forcing on climate timescales in the Pacific (Fig. 1) organize the response and phase relationships among the ecosystem components over similarly large scales, although the local response of the ecosystem is strongly dependent on regional conditions.

3.1 The North Pacific Ocean

As the Aleutian Low varies on decadal timescales, two characteristic channels of large-scale oceanic response emerge: the canonical SST pattern driven directly by the surface atmospheric forcing fields; and the KOE SST pattern associated with gyre-scale circulation and thermocline changes (Miller and Schneider, 2000). These two channels of response organize biological activity in the North Pacific Ocean in a number of ways.

The canonical SST pattern links changes in the California Current and in the Gulf of Alaska to those in the central North Pacific and the equatorial region (where part of the SST forcing originates). The local mechanisms by which these North Pacific and equatorial Pacific regions respond to changes in SST, and the concomitant changes mixed-layer depth (MLD), thermocline stability and horizontal currents, can be subtle and are explained in subsequent sections. As an example of how this organization occurs, Fig. 2 (top panels) shows the relationship between the canonical SST pattern and ecosystem response in the simulation by Chai *et al.* (2003a). Forced by 1960–1998 winds, the physical ocean model drives the ecosystem model. The principal component time series for the combined physical-biological empirical orthogonal function (EOF) shows the transition to cold central N Pacific conditions in the mid-1970s and a return to near neutral conditions after the late 1980s. The largest biological response is around the subtropical front, and a coherent large-scale biological response extends northeastwards along the western North American coast and into the Gulf of Alaska.

The KOE SST pattern occurs with roughly a two-to-five year lag after the canonical pattern is established and thus provides a lagged forcing for the biological fields of that region of the subarctic front. Since it is driven by upwelling and horizontal current variations from incoming Rossby waves, the mechanisms by which the ecosystem responds are again distinct from the canonical SST surface forced effects. Wind-stress curl changes that influence the Kuroshio in later years will also force a coherent response in the Gulf of Alaska (Lagerloef, 1995), altering the mean currents and eddy fields there, that is closer to being in phase with the canonical SST pattern because open-ocean Rossby waves are not important at those high latitudes (Fu and Qiu, 2002; Cummins and Lagerloef, 2003). As an example of how this organization occurs, Fig. 2 (bottom panels) shows the relationship between thermocline variability and the ecosystem response in the simulation by Chai *et al.* (2003a). The principal component time series shows the smooth transition to western-intensified cold conditions north of the KOE (Miller *et al.*, 1998) in the 1980s and a switch to warm conditions after the late 1980s. Although the largest biological response is around the subtropical front, a coherent large-scale biological response occurs north of Japan and in the KOE region in response to physical oceanographic changes. The thermocline changes seen here are clearly spatially matched to the biological changes around Japan and in the KOE region (unlike the SST there). Note, however, that the strong biological response around the subtropical front appears in this EOF because it is temporally correlated to the thermocline changes even though it is physically driven by changes in surface mixed layer conditions (Polovina *et al.*, 1995). Note also that

the model thermocline (and local SST) response in the KOE region does not exhibit a strong lag with respect to the canonical SST. The KOE region of this particular model seems to be dominated by a fast barotropic response and does not admit baroclinic Rossby waves in this latitude band, a result that is currently under investigation.

Changes in fresh-water fluxes are also correlated to large-scale climate variations through coastal streamflow alterations and open-ocean variations in precipitation and evaporation. Streamflow changes are known to have a profound influence on the surface mixed layer and the coastal circulation in the Alaska Current, the Alaska Stream and the Alaska Coastal Current (e.g., Royer *et al.*, 2001). The structure of the open-ocean mixed layer is strongly linked to salinity in the northeast Pacific (Freeland *et al.*, 1997; Haug *et al.*, 1999). Hence fresh-water fluxes exert a strong control on nutrient fluxes from the deep ocean to the euphotic zone in these subarctic regions and result in the biological activity being structured with climate variations.

3.2 The Tropical Pacific Ocean

Decadal climate variability of the tropical Pacific is much weaker than the dominant climate variability due to ENSO, which occurs with periods of two-to-seven years (Philander, 1999). Fiedler (2002) summarized how biological effects, from phytoplankton productivity to seabirds and penguins, are organized in the tropical Pacific by ENSO variability. During El Niño, reduced nutrient flux dramatically reduces productivity and disrupts the food web across the equatorial and coastal environments of the eastern tropical Pacific (Barber and Chavez, 1983; Chavez *et al.*, 1999). ENSO impacts on the lower trophic ecosystem are largely due to changes in nutrient flux to the euphotic zone (Fiedler *et al.*, 1992; Barber *et al.*, 1996). Nutrient flux is a function of the nutrient and vertical velocity that is driven primarily by the zonal surface wind. In the central and eastern equatorial Pacific the location of nutrient source waters is correlated and well defined by the depth of the 20°C isotherm that is located in the center of the thermocline (Le Borgne *et al.*, 2002). The equatorial undercurrent flows from west to east across the basin at the depth of the thermocline and is the nutrient source for the equatorial Pacific pelagic ecosystem (Toggweiler and Carson, 1995; Jiang *et al.*, 2003). On average, the 20°C isotherm slopes upward more than 150 m from the western to the eastern equatorial Pacific (McPhaden *et al.*, 1998). During El Niño, depression of the thermocline in the eastern Pacific and elevation of the thermocline in the western Pacific causes the thermocline to level out, thereby reducing the upward nutrient fluxes.

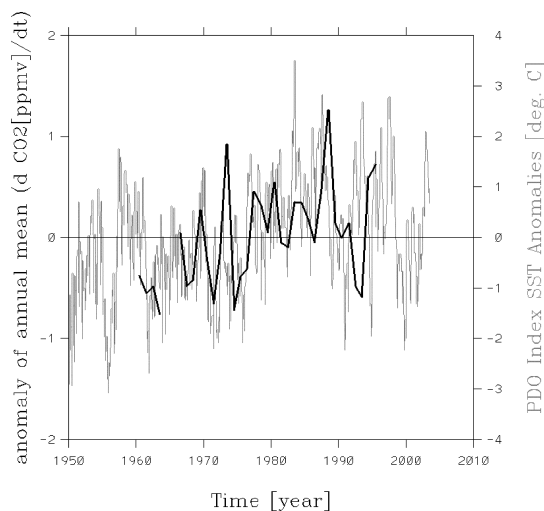The equatorial Pacific Ocean also exhibits variations at decadal time scales, and the changing Pacific climate

Fig. 3. Anomaly of the annual tendency of CO_2 (dark line) observed at Moana Loa compared to the PDO index (grey line). The CO_2 increase and changes observed in the Pacific is linked to the physical climate decadal variability.

has direct impacts on marine ecosystem and carbon cycle (Fiedler, 2002). However, the dynamic linkages between ocean physics and biology at the decadal scales are poorly understood because it is difficult to perform oceanographic sampling at the time and space scales necessary to elucidate the connections between large-scale physical processes and the smaller-scale biological responses (McGowan *et al.*, 1998). For example, in the Pacific, air and ocean temperatures, fish landings and ocean productivity show synchronous variations on multi-decadal time scales (Chavez *et al.*, 2003), but there have been no well articulated hypotheses to link the environmental variables with ecosystem responses.

Recent studies of decadal changes in ocean circulation in the tropical and subtropical Pacific by McPhaden and Zhang (2002) have demonstrated a slowdown in the meridional overturning in the upper Pacific Ocean. They observed a 25% decrease in the transport convergences and a commensurate 0.8°C increase in tropical surface waters over the past 25 years that corresponded with the 1976–77 climate shift (Miller *et al.*, 1994b; Trenberth and Hurrell, 1994). Decreasing in the equatorial upwelling since 1976–77 would result in lowering nutrients and CO_2 supply to the surface, therefore, decreasing in primary productivity and reduction in the sea-to-air CO_2 from the equatorial Pacific Ocean.

The impacts of decadal scale climate variability in the equatorial Pacific on biogeochemistry and ecosystems are poorly sampled and the limited observational evidence is not conclusive (Fielder, 2002). Some limited physical evidence and modeling results have shown the decadal

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variations modulate the El Niño intensity and frequency in the equatorial Pacific (Latif *et al*., 1997; Fedorov and Philander, 2000). When the PDO index and ENSO are both in the warm phase, as happened during the 1997–98 ENSO event, the SST anomalies can be extremely large along the equatorial Pacific and the eastern side of the Pacific Basin. During the strong 1997–98 El Niño, the surface nutrients and chlorophyll diminished and the equatorial Pacific Ocean retained 1.0 PgC of carbon that normally would have been lost to the atmosphere as carbon dioxide (Chavez *et al*., 1999). There have been some indications that a regime shift in the Pacific has occurred just prior to or after the 1997–98 El Niño (Schwing and Moore, 2000; Minobe, 2000; Schwing *et al*., 2002a). The alteration of the equatorial Pacific biogeochemistry and marine ecosystem in the new climate regime needs to be investigated much more extensively since the rate of change of CO_2 observed on Moana Loa is clearly correlated to the PDO index (Fig. 3).

4. Regional Pacific Oceanic Ecosystem Variability Organized by Climate

Because of the large-scale organization of biological response in the Pacific by physical climate forcing, distinct regional patterns arise in the ecosystem due to regional differences in other variables besides SST and thermocline structure. The main areas of interest here are the tropical Pacific, the central North Pacific, the Kuroshio/Oyashio Extension, the Gulf of Alaska and the California Current System. While many other regions could be considered as well, we focus on these as being clearly linked the large-scale physical forcing and because they have been studied the most thoroughly.

4.1 Tropical Pacific ecosystem variations

Equatorial Pacific physical climate variations strongly influence not only local but also global biogeochemistry. In the central and eastern equatorial Pacific Ocean, the easterly Trade Winds blowing along the equator create a divergence in the surface flow field that generates the cold tongue of upwelled water which spans approximately one quarter of Earth's circumference. The biogeochemical consequences of equatorial upwelling are high concentrations of inorganic nutrients (nitrate and phosphate) in the euphotic zone, relatively elevated phytoplankton biomass, and high concentrations of total inorganic carbon (TCO_2) that results in CO_2 outgassing from the ocean surface (Murray *et al*., 1994; Feely *et al*., 2002). The flux of nutrients to surface waters is primarily governed by upwelling rates, depth of the thermocline and the dynamics and chemical composition of the equatorial undercurrent (Chai *et al*., 1996; Gordon *et al*., 1997).

Due to the zonal extension of the equatorial Pacific Ocean (over 200° of longitude), there is lack of sufficient

eolian dust deposition to the equatorial Pacific (Duce and Tindale, 1991), as well as lack of iron transport from the continental margins to the open equatorial regions (Mackey *et al*., 2002). The iron deficiency maintains high nitrate low chlorophyll (HNLC) conditions in the equatorial Pacific, and two iron addition experiments have highlighted the important role of iron in equatorial HNLC waters (Martin *et al*., 1994; Coale *et al*., 1996). The large portion of phytoplankton in the equatorial HNLC waters is grazed by zooplankton, and this grazing is one of the most important processes driving export of POC to the deep sea. JGOFS equatorial Pacific studies provided convincing evidence that microzooplankton ($<{\sim}200\ \mu$m) are the main grazers consuming 65–70% of primary production under normal upwelling conditions, Landry *et al*. (1997) concluded that HNLC conditions in the central and eastern equatorial Pacific result from the interacting effects of Fe limitation and microzooplankton grazing. Beside iron limitation and grazing control, there has been some evidence for silicate regulation of new production during periods where diatoms dominate the phytoplankton community in the central and eastern equatorial Pacific (Dugdale and Wilkerson, 1998), and some of the mechanisms controlling plankton community structure have been explored mathematically by adding the Si cycle and the diatoms' Si requirement to biogeochemical models of the upper ocean (Dugdale *et al*., 1995; Chai *et al*., 2002).

During the 1997–98 El Niño, deepening of the thermocline and cessation of upwelling forced nutrient depletion through the upper 100 m and impoverishment of surface chlorophyll (Strutton and Chavez, 2000). In contrast, during the ensuing La Niña, the thermocline was shallow across the basin, and the upwelling field was strongly developed due to the intensifying of the trade wind. Shoaling of nutrient source waters into the region of maximum upwelling velocity greatly increased nutrient flux to the ecosystem, as evidenced by the extreme bloom that occurred during summer 1998 (Chavez *et al*., 1999).

It is well known that equatorial Pacific climate variability (mostly associated with ENSO) affects Peruvian and Chilean fisheries, particularly those of the small pelagic fisheries. During El Niño, reproduction and even survival of Peruvian anchovy are seriously compromised due to reduced food supplies for adults and larvae, and yet warm water species like shrimp and sardines moved southward from the equator (Barber and Chavez, 1986). ENSO has effects on commercially exploited fish as well. For example, the abundance fluctuations, distributions and migrations of tropical tuna species (e.g., skipjack, yellowfin, bigeye, albacore) have been shown to be sensitive to environmental variability (Lehodey *et al*., 1997). In particular, El Niño/La Niña and decadal climate variability appear to have important consequences for spatial
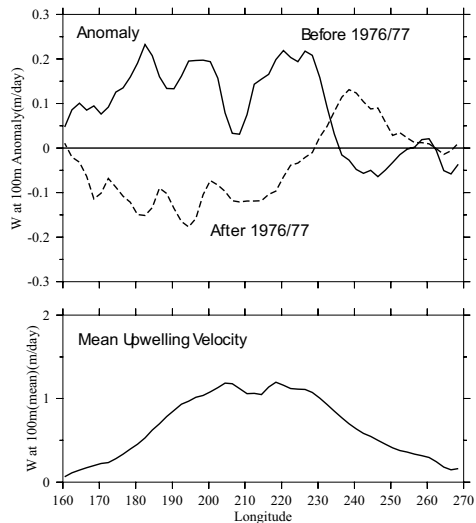
Fig. 4. (Top) Modeled upwelling velocity anomaly at 100 m depth along the equator averaged between 2.5°N and 2.5°S. (Bottom) Modeled mean upwelling velocity (m/day) at 100 m depth along the equator averaged between 2.5°N and 2.5°S. The dashed line shows the weaker upwelling (negative anomaly) after the 1976–77 climate shift in the Pacific, the solid line shows stronger upwelling (positive anomaly) before 1976/77. During the positive upwelling period, the equatorial Pacific tends to be more productive due to more supply of nutrients to the euphotic zone. The concomitant decadal SST anomaly created atmospheric teleconnections that strengthened the Aleutian Low after the shift and caused basin-scale changes in North Pacific productivity correlated to these equatorial changes. Model characteristics are discussed by Chai *et al*. (2003a).

distributions and migrations of the tuna populations. While skipjack and yellowfin had higher (lower) recruitments during El Niño (La Niña) events, the subtropical albacore species showed the opposite pattern, with lower (higher) recruitment during El Niño (La Niña) (Lehodey *et al*., 2003). Production of higher trophic levels (usually exploited species) depends on the production at lower levels (bottom-up control) and may be modulated by the physical forcing and the structure of marine ecosystems.

Long-term observations of biological and chemical variables are not available for the central and eastern equatorial Pacific. A few survey cruises concentrated in the equatorial Pacific show a strong correlation between surface nutrient content and chlorophyll, that both variables showed a decreasing trend slowly during the 1980s and early 90s (Chavez *et al*., 2003). Combining the NODC World Ocean Database and available satellite ocean color data (CZCS and SeaWiFS), Gregg and Conkright (2002) have found the surface chlorophyll concentration decreased slightly in eastern equatorial Pacific since 1980s,

but there were no changes in the central equatorial Pacific.

Using a three-dimensional physical-biogeochemical model forced with COADS wind stress and heat flux, Chai *et al*. (2003b) have investigated biogeochemical responses to the decadal physical variability in the equatorial Pacific Ocean. The circulation model resolved the decadal variations between 1950s and 1990s are consistent with the observed variability in the tropical Pacific Ocean reported by McPhaden and Zhang (2002). Their biogeochemical model (Chai *et al*., 2003b) also captured the responses to the changing physical conditions in the equatorial Pacific with an overall decrease in the equatorial upwelling transports of silicate and nitrate, and a 10% decrease in the modeled primary production and phytoplankton biomass since 1970s (Fig. 4).

4.2 Central North Pacific ecosystem variations

In the central North Pacific, a two-fold increase in the integrated chlorophyll-a was observed in the pelagic ecosystem during summer (Venrick *et al*., 1987) that was clearly linked to the changed oceanographic conditions (deeper mixed layer and colder SST) there after the 1976–77 climate shift which persisted for at least a decade. In the northwestern subtropical gyre region, on the other hand, chlorophyll a in spring showed a steady increase from the mid 1970s to the mid 1980s (Limsakul *et al*., 2001).

Based upon the Hawaii Ocean Time-series (HOT) program and the limited historical data, Karl *et al*. (2001) have shown that primary productivity in the north Pacific subtropical gyre was significantly lower before the mid 1970s than it was in the early years of the (HOT) program. They also suggest a shift in community structure with increases in nitrogen-fixing organisms. During the first decade of the HOT program the concentration of dissolved phosphorus in the surface ocean also slowly declined. The source of this phosphorus is likely from eastern and equatorial Pacific upwelling. The decrease, the HOT researchers surmise, was due to an increase in nitrogen fixation.

Analyzing and comparing apparent oxygen utilization (AOU) from four meridional transects through the northeast subtropical Pacific between 1980 and 1997, Emerson *et al*. (2001) found an increase in AOU by 20–25% over the past two decades and suggested an increase in oxygen demand by the biological pump and/or oxygen degassing by upper-ocean ventilation were accountable for the observed AOU increase.

Analyzing the estimated winter and spring MLD from the National Oceanographic Data Center (NODC) temperature data, Polovina *et al*. (1995) found that winter and spring MLD in the subtropical and transition zone was 30–80% greater during 1977–88 than during 1960–

76, while in the subarctic zone the MLD was 20–30% shallower. They also employed a simple plankton model forced with observed MLD and nutrients, and found that deepening in mixed layer might increase phytoplankton production in nutrient-poor subtropical regions by supplying more nutrients from depth, shallowing MLD in the subarctic would also increase production by increasing light availability to phytoplankton cells within the mixed layer. In between the subtropical and subarctic regions, both nutrient supply and light availability likely co-limit phytoplankton production, but how these processes change on the decadal time scale is not clear.

Chai *et al*. (2003a) used a physical-biological model to simulate the response of the Pacific ecosystems to the Pacific decadal climate variability. They found that the modeled winter MLD shows the largest increase between 30°N and 40°N in the central North Pacific, with a value of 40–60% higher during 1979–90 relative to 1964–75 values. The winter Ekman pumping velocity difference between 1979–90 and 1964–75 shows the largest increase located between 30°N and 45°N in the central and eastern North Pacific. Due to these changes in the mixed layer depth and Ekman pumping velocity, the modeled winter surface nitrate concentration also showed an increase after 1976–77 climate shift in the latitudinal band between 30°N and 45°N from the west to the east (135°E–135°W). The modeled nitrate concentration increases after 1976–77 enhances primary productivity in the central North Pacific. Enhanced primary productivity after the 1976–77 climate shift contributes higher phytoplankton biomass and therefore elevates chlorophyll level in the central North Pacific. Increase in the modeled chlorophyll expand the chlorophyll transitional zone and push the TZCF equatorward (Chai *et al*., 2003a).

The central gyre of the Pacific Ocean has a relatively weak anticyclonic circulation pattern, effectively isolates the upper portion of the water column from large volume water exchange with the surrounding waters (Karl and Lukas, 1996). Consequently, the central gyre of the Pacific Ocean is characterized as relatively homogeneous in horizontal with a deep permanent pycnocline and nutricline (Hayward, 1987). Seasonal changes of the upper water properties, including mixed layer depth, are also weak near the Hawaii Ocean Time-series (HOT) station (Bingham and Lukas, 1996). On longer time scales, both El Niño events and decadal variability in physical processes influence both macro- and micro-nutrient concentrations, phytoplankton biomass, primary productivity, particulate C:P and C:N ratios, and food web structure (Venrick *et al*., 1987; Karl, 1999). Even after more than 10 years of time-series observations at the HOT, the linkages between the underline physical processes and changing biological and chemical cycles are still not clear (Karl *et al*., 2001).

4.3 Kuroshio-Oyashio Extension ecosystem variations

The Kuroshio current and Oyashio current extension areas are sometimes classified as a single regional entity (KOE region) around the western boundary current region, because of its importance in studies of large-scale ocean-atmosphere interaction (Miller and Schneider, 2000; Schneider and Miller, 2001; Qiu, 2002; Yasuda, 2003). The KOE has significant role in terms of mechanisms of oceanic response to the climate change and in possible feedback to the atmospheric condition. In terms of ecological variation, however, it is preferable to separate the Kuroshio and the Oyashio as in the North Pacific ecosystem model by Kawamiya *et al*. (2000) because food web structure, biological processes and ecosystem response to common climatic and physical forcing differ considerably between the two regions. Furthermore, the Oyashio region, located on the western edge of the western subarctic gyre, is known for high new production (Saito *et al*., 2002) and high biological drawdown of pCO$_2$ (Takahashi *et al*., 2002). Therefore, elucidating the response of the Oyashio ecosystem to climate variability is of particular interest of biogeochemical studies. Figure 5 shows a schematic diagram of these regional KOE ecosystem responses and their links to climate forcing.

In Oyashio waters, wintertime mixing richly supplies the surface waters with nutrients and micronutrients including iron (Ono *et al*., 2002) from subsurface levels. This causes extensive diatom blooms to occur in spring when the sea surface warms to form a stable, shallow mixed layer. Dominant secondary producers are large copepods with a relatively long life cycle, which perform ontogenetic vertical migration. Material transport is efficient through this relatively simple grazing food chain (Taniguchi, 1999; Saito *et al*., 2002) and the active transport by migrating animals (Kobari *et al*., 2003). In contrast, biological production is nutrient-limited in the Kuroshio region and in the northwestern subtropical gyre where thermal stratification develops even in winter (Sugimoto and Tadokoro, 1998). Phytoplankton production is highest in February and soon diminishes as nutrient depletion occurs (Limsakul *et al*., 2002). Dominant secondary producers are smaller copepods with a relatively short life cycle (Nakata *et al*., 2001), and the relative importance of the microbial loop must increase as found for other subtropical waters (e.g., Karl *et al*., 2001).

The intensification of the Aleutian Low enhanced the wintertime East Asian Monsoon (westerly winds) over Japanese waters after the 1976–77 climate shift through 1988 (Limsakul *et al*., 2001). However, lower trophic level ecosystems responded to the climatic forcing in different ways in the Oyashio and Kuroshio regions. Spring biomass of phytoplankton (Ono *et al*., 2002) and zooplankton (Kubo, 1994; Odate, 1994; Tomosada and Odate, 1995; Sugimoto and Tadokoro, 1997) was reduced
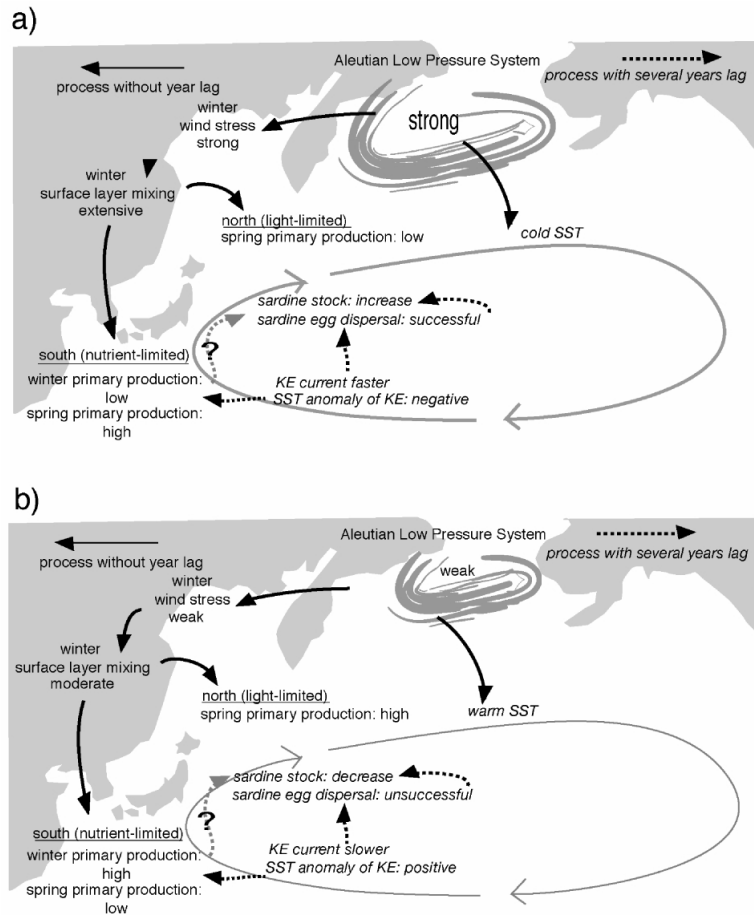
Fig. 5. Schematic diagram of decadal scale ecosystem changes in the KOE region for (a) strong Aleutian Low (positive PDO index) and (b) weak Aleutian Low (negative PDO index). Processes occurring in phase with the Aleutian Low are shown in normal letters with solid arrows. Processes occurring with a several-year lag are shown in italics and with broken arrows.

during this period in the Oyashio region, but increased south of the Kuroshio region (Limsakul *et al.*, 2001). These results are consistent with the theory demonstrated by Polovina *et al.* (1995) that strong winter mixing affects lower trophic level productivity negatively in light-limiting subarctic water and positively in nutrient-limiting subtropical water. Another example of this effect was observed in the north-south contrast of multi-decadal scale variability of plankton abundance in the Japan Sea, which contains branches of both the subarctic and subtropical gyres in its semi-closed basin (Chiba, in preparation, 2003). We must note that the effects of long-term changes in wintertime vertical mixing could induce quantitative variation as well as phenological changes in ecosystem processes. For example, after the shift of the mid-1970's, the phytoplankton biomass south of the Kuroshio decreased in wintertime but increased in springtime (Limsakul *et al.*, 2001). This suggests that the timing of the re-stratification of surface waters, favorable for

phytoplankton growth, was delayed by strong winter mixing.

In addition to direct atmospheric forcing, such as wind stress and surface heat fluxes driving local mixing changes, associated with the PDO index, oceanic forcing associated with Rossby waves derived by the past PDO signal is significantly responsible for SST anomaly in the KOE region (Schneider and Miller, 2001). The effects of these Rossby waves on the ecosystem has not been determined, although Fig. 2 (bottom panels) shows that a raised model thermocline in this region produces increased biological productivity in an ecosystem model, and vice versa for a depressed thermocline. Kuroshio meandering was strong and its flux increased after the mid 1970s (Nakamura and Hinata, 1999; Qiu and Miao, 2000) corresponding to an intensification of westerly winds and the western subtropical gyre (Hanawa, 1991). Furthermore, since the Kuroshio plankton communities with relatively short life cycles tend to be sensitive to changes in

ambient conditions (Nakata *et al*., 2001), the atmospheric and hydrographic changes caused a complex lower trophic level response. Besides the decadal changes in the Oyashio region that are related to the Aleutian Low variations, long-term linear trends were reported in the hydrographic, chemical and biological properties as well (Ono *et al*., 2001, 2002; Chiba *et al*., 2004). It is essential to distinguish the mechanisms controlling, and the possible interactions between, the observed ecosystem responses to the decadal-scale changes and the long-term trend. Variation of higher trophic level production might be more influenced by advective transport changes (e.g., Qiu and Joyce, 1992; Deser *et al*., 1999; Hanawa and Kamada, 2001) rather than direct atmospheric forcing as discussed later.

4.4 Bering Sea ecosystem variations

The Bering Sea exhibits a range of physical and biological responses to atmospheric forcing on decadal scales. There has been a change in sea ice over the last few decades towards faster growth in winter and an earlier spring thaw (Stabeno and Overland, 2001; Stabeno *et al*., 2001). Besides affecting the seasonality of the regional CO_2 budget (Pipko *et al*., 2002), the earlier spring thaw has also caused a shift towards earlier spring phytoplankton blooms. Additionally, since 1997 there have been blooms of coccolithophores each summer (Vance *et al*., 1998). Evidence suggests that these blooms, which appear despite supposedly poor conditions for their growth, are sustained by mixing processes similar to the transport of cold slope water via eddies into the surface waters (Shin *et al*., 2002). This transport is, in turn, affected by the annually variable strength of eddies and circulation in the Bering Sea, and related, possibly, to the interannual variability of sea surface temperature (Iida *et al*., 2002). Weaker blooms accompanied cold-water years, an observation mirrored by Coyle and Pinchuk (2002) who found that copepod production during warmer years (which have higher phytoplankton production) was higher than during colder years.

In the higher-order trophic levels, direct and indirect influences of climate changes have been detected. For example, changes in the distribution of salmon were recorded that were related to the 1976–1977 and 1989 regime shifts in the North Pacific (Ishida *et al*., 2002). Shifts have been reported in populations of crab, seabirds, and marine mammals as well (Macklin *et al*., 2002). Conversely, walleye pollock distributions were found to vary with sea ice fluctuations (Hunt *et al*., 2002).

4.5 Gulf of Alaska ecosystem variations

Changes in climate in the northeast Pacific have been linked to many aspects of ecosystem variations in the Gulf of Alaska (e.g., Hayward, 1997; Francis *et al*., 1998;

Miller and Schneider, 2000; Benson and Trites, 2002). These atmospheric climate changes occurred at various timescales, from annual through decadal, and were accompanied by physical oceanographic responses with similar time scales. Ecosystems also respond on similar time scales with tighter coupling to physical changes occurring at the lower trophic levels.

In the absence of consistent long-term measurements of phytoplankton and zooplankton abundance, Benson and Trites (2002) concluded that shifts in biological production generally correspond to the timing of climate shifts. At the time of the 1976–77 climate shift, the mixed layer in the Gulf of Alaska shoaled 20–30% (at least based on a temperature criterion; see Freeland *et al*. (1997), for discussion of salinity effects), relative to 1960–1976. Modeling studies indicated an increase in phytoplankton as a result of increased light levels across the shallower mixed layer (Polovina *et al*., 1995). Sugimoto and Tadokoro (1997) found that chlorophyll concentrations in the eastern Bering Sea and eastern Subartic Pacific increased in the mid-1960s, peaking during the mid-1960s to early 1970s in the central and western subarctic Pacific, and remaining high across the subarctic Pacific until the end of the 1980s after which they declined.

Zooplankton biomass responded to the 1976–77 climate shift also by increasing significantly across the subarctic North Pacific from 1956–1962 to 1980–1989 (Brodeur and Ware, 1992; Brodeur *et al*., 1999; Mackas *et al*., 2001; Benson and Trites, 2002). As with chlorophyll, Sugimoto and Tadokoro (1997) found that zooplankton biomass in the Bering Sea and subarctic Pacific started increasing in the mid-1960s remaining high until the late 1980s. Summer zooplankton biomass doubled between the late 1950s and 1980s in the Central Gulf of Alaska (Brodeur *et al*., 1996). Wickett (1967), studying interannual zooplankton volumes during the 1950s and early 1960s found implications that zooplankton volume, and nutrients, carried in the subarctic frontal currents enrich areas downstream of the bifurcation of the West Wind Drift into the California and Alaska Currents. Changes in surface winds that favor a strengthening of either downstream current will preferentially enrich that current. Strub and James (2003) find evidence for this effect in their analysis of satellite sea level measurements. While more nuanced mechanisms have been put forward (e.g., Roemmich and McGowan, 1995) this differential enrichment, in general, is supported by the observation that zooplankton biomasses in the Gulf of Alaska and the California Current appear to have responded in opposite directions to the 1976–77 climate shift (Brodeur and Ware, 1992; Roemmich and McGowan, 1995; Brodeur *et al*., 1996). However, warming of the upper ocean in the California Current suppresses upwelling of nutrients that would then lead to reduced productivity, which is an al-

ternative explanation for the opposite effect being found in the light-limited Gulf of Alaska (McGowan *et al.*, 2003; Di Lorenzo *et al.*, 2003).

Fish species in the Gulf of Alaska may respond on many timescales in the coupled atmosphere/ocean (Francis *et al.*, 1998). Power spectra for anchovies and sardines, in the California Current, show peaks at around 60 and 75 years which correspond to cold and warm periods that occur every 50 to 75 years (Baumgartner *et al.*, 1992; Minobe, 2000). Latif and Barnett (1994) identified an approximately 20 year oscillation in the North Pacific SST. The 6-to-12 year warm/cool eras reported by Hollowed and Wooster (1992, 1995), for records between 1932–1988, were linked to strong year classes of groundfish from California to the Bering Sea although the relationship between climate and stock abundance may be related to the "storage effect" found in strong year classes. Warm temperatures are not sufficient, in any case, to produce a strong year class of groundfish.

Herring also exhibit a positive relationship to SST in the Sitka Sound between 1972 and 1990 (Zebdi and Collie, 1995). They suggested that warmer temperatures led to earlier spawning which provided better conditions for survival although, as Francis *et al.* (1998) point out, this period overlapped with the increase in zooplankton biomass reported by Brodeur and Ware (1992) in the Gulf of Alaska. Interestingly, Zebdi and Collie (1995) reported that northern and southern stocks, out of phase by 1 year, reacted oppositely to the temperature changes. Northern Stocks responded positively to warmer temperatures while Southern Stocks responded negatively. This, however, may be related to climate conditions enhancing the northward flow of zooplankton and nutrient laden warm water from the subarctic frontal currents.

Salmon stock variations along the entire North American west coast are increasingly being linked to decadal climate changes (Beamish and Bouillon, 1993; Mantua *et al.*, 1997). Stocks in Alaska tend to vary in phase with each other, but out of phase with northwestern U.S. on stocks (Francis and Sibley, 1991; Hare *et al.*, 1999). The mechanisms for this are still unclear (Hare and Francis, 1995; Benson and Trites, 2002) but early life history effects on populations (Beamish and Bouillon, 1993) and the north-south asymmetry of production in response to the canonical SST pattern are the prevailing theories.

4.6 California Current System ecosystem variations

The California Current Ecosystem is one of the most studied Eastern Boundary Current (EBC) ecosystems in the world. Within this region, one of the longest running observational efforts, the California Cooperative Ocean Fisheries Investigation (CalCOFI), has been collecting data in the Southern California region since the 1950s.

True to its nature of being an EBC, the large-scale circulation pattern is dominated by the California Current, an equatorward meandering surface current that is obscured by the high eddy kinetic energy generated by a rich coastal eddy field (Swenson and Niiler, 1996; Miller *et al.*, 1999), and the poleward flowing California Undercurrent. Superimposed upon these circulation features is an alongshore wind-driven coastal upwelling process that supports a highly productive ecosystem.

Satellite observations in the early 1980s of sea surface temperature and ocean chlorophyll *a* demonstrated the presence of filament-like structures propagating offshore during periods of intense coastal upwelling—most active in the spring and summer months (Brink and Cowles, 1991; Strub *et al.*, 1991). The variability of the satellite-derived chlorophyll field in the California Current was characterized by Strub *et al.* (1990) who noted that the largest nonseasonal anomaly in the observed chlorophyll fields was related to the large-scale decrease during the 1982–1983 El Niño with its effect being stronger and longer lasting in the southern regions of the California Current System.

Roemmich (1992) first noted increases in upper ocean temperatures and sea level along the California coast by analyzing the CalCOFI temperature data from 1950 to 1991. Further analysis by Roemmich and McGowan (1995) on the CalCOFI zooplankton biomass data set observed that during within this time period of the CalCOFI observations (1950-present) the surface waters in the CalCOFI region underwent a warming trend while macrozooplankton biomass decreased by 70%. In conjunction with the increase in ocean temperatures was an observed deepening of the main thermocline which McGowan *et al.* (2003) argue supported a decline in primary production that lead to the observed large scale decline in the zooplankton biomass. Numerical experiments by Di Lorenzo *et al.* (2003) show that upwelling of nutrient rich waters are indeed suppressed due to warming induced by surface heat fluxes, in spite of a long-term increase in upwelling favorable winds (Schwing and Mendelsohn, 1997).

After seasonal variability, ENSO oscillations are the most significant source of variability in the California Current System. While ENSO perturbations are short lived and are primarily centered on the equatorial regions, they can influence ecosystems of the North Pacific (Wooster and Fluharty, 1986; Mysak, 1986; Chavez, 1996; Bograd and Lynn, 2001). On longer time scales, the PDO pattern is markedly similar in spatial scale to ENSO, having large positive SST anomalies over the Pacific equatorial regions during the warm phase with associated cooling in the northwestern Pacific (Zhang *et al.*, 1997). While it has been argued that many of the potential ecosystem impacts have yet to be clearly demonstrated (Miller and

Schneider, 2000), some simple links have been argued as a control between the physical processes and the observed changes to the ecosystem (Mantua *et al*., 1997; Chavez *et al*., 2003). The long-term decline in the zooplankton in the California Current System is argued to be due to a deepening of the thermocline with an associated warming and increased mixed layer depth from increased wind stirring (McGowan *et al*., 2003).

While the increase in gyre circulation in the subtropics is argued to lead to an increase in production in the centers of both gyres (Karl *et al*., 2001) as well as the edge of the subarctic gyre, the effects are reversed along the California coasts, where the strength of the California Current and California Undercurrent and the magnitude of the coastal upwelling are diminished. The response of upwelling to these changes seems to result from deeper thermoclines as upwelling estimates during the warm phase of the PDO index support an increase in upwelling during this period (Schwing and Mendelsohn, 1997).

The increase in ocean temperatures along the CCS is part of the basin-scale PDO index. Associated with the PDO index are variations in the Aleutian Low (Miller *et al*., 1994b) which controls the strength of the alongshore winds and coastal upwelling and modifies the heat budgets. During the period between 1960 and 1990 the CCS was transitioning between the cold (more productive) to warm (less productive) phase of the PDO index. During the cold phase of the PDO index, the Aleutian Low winds are increased along the CCS and support stronger periods of upwelling. During the warm phase, the coastal winds are reduced and coastal upwelling subsides. As a result, the CCS experiences a decrease (20–30%) in productivity during the transition between the cold to warm phases of the PDO index. This decrease in coastal productivity has reportedly been the recent cause of zooplankton biomass. While California Current region temperature variability does parallel that observed in the subtropical gyre (Miller *et al*., 1994a, b; Polovina *et al*., 1995), it is not a nutrient-limited region of the ocean and therefore deeper MLDs should not enhance productivity but rather decrease it due to increased light limitation. Recent results from Di Lorenzo *et al*. (2003) demonstrate a complex change occurred during the recent decadal warming event in the CCS. Local heat flux anomalies caused an increase in the upper ocean temperature and led to a decrease in the isopycnals/thermocline. This drop in thermocline removed nutrients away from the upwelling zone which was increasing as a result of increased alongshore winds. Di Lorenzo *et al*. (2003) argue that the decrease in thermocline depth was more effective at reducing the nutrient flux to the upper ocean than the observed increase in upwelling.

While the CalCOFI program has provided a rich time series of observations that allow researchers to analyze the decadal variability in the last half of the past century, a number of sedimentary basins (23) exist within the Southern California Bight (Emery, 1960) that have yielded additional information on the inter-decadal signals during the last several millennium using varve records from the Santa Barbara basin (Baumgartner *et al*., 1992; Biondi *et al*., 1997). Both of these data sets reveal that the California Current region undergoes decadal-scale variability. The longer time series analysis presented by Biondi *et al*. (1997) identified four oscillations in varve thickness at periods of ~12, ~25, ~58, and ~100 years. In addition, since around 1600 AD, the laminated structure of the sediments have been preserved, while prior to that period it was destroyed through bioturbation processes. Lamination in sediments is used as a proxy for anoxic conditions in the sedimentary basins. It has been argued that this change in preservation mode is a reflection of a larger scale change in the circulation patterns of the California coast due changes in atmospheric forcing (Lange *et al*., 1990). More recent observations (Stott *et al*., 2000) have linked an increase in Pacific intermediate water dissolved oxygen concentrations with a return to bioturbated (non-laminated) sediments and a decrease in carbon oxidation rates in the sediments due to a decrease in the level of organic matter sedimentation. This decrease in carbon sedimentation rates is supported by the reduced level of upwelling (20–30%)—which controls the total supply of new production—and the increase in sea temperatures-which controls the rate of water column organic mater remineralization.

5. Mechanisms of Interaction

We next describe the specific mechanisms by which ocean biological variables are likely to interact with the large-scale organized patterns of oceanic climate forcing. The success of ocean ecosystems depends upon the availability of light, temperature, nutrients, space, and the timing at which they are made available. Climate variations influence the ocean biology through changing the solar flux, temperature, horizontal current advection, vertical mixing and upwelling, fresh-water fluxes and sea ice. Intrinsic biological interactions, in the presence of stationary climate forcing, are also important.

Oceanic biological processes also modify the flow of energy through the ocean-atmospheric climate system in two major ways (Miller *et al*., 2003). The first is through the effect of phytoplankton on upper-ocean absorption of solar radiation (Nakamoto *et al*., 2001; Murtugudde *et al*., 2002; Strutton and Chavez, 2003). The second is through the flux of dimethylsulfide (DMS) to the atmosphere, which affects cloud formation (Charlson *et al*., 1987). The two hot spots in the Pacific where these effects might wield their greatest influence are in the tropical Pacific cold tongue region (where SST and atmos-

pheric teleconnections are sensitively linked) and in the KOE region (where midlatitude ocean-atmosphere feedbacks are likely to be strongest and most sensitive).

5.1 Solar flux effects on ecosystems

Solar energy is the primary driving force for all processes in the ocean. The magnitude of the flux of solar energy has been demonstrated to covary with the sunspot cycle (Eddy, 1977). In addition, cloud levels, which further reduces the amount of Photosynthetically Available Radiance (PAR) for phytoplankton growth is also linked solar variability (Svensmark and Friis-Christensen, 1997). Cloud cover also increases the relative role of phytoplankton in modulating the attenuation scale of solar radiation into the ocean (Siegel et al., 1999).

5.2 Temperature effects on ecosystems

Temperature controls the metabolic rates of phytoplankton, zooplankton and bacteria. Phytoplankton maximum growth rates are limited by temperature. Individual phytoplankton species, however, each have an optimal temperature for growth below or above which they will grow at a slower rate. As a result of this and the variability of temperature in the open ocean, phytoplankton are typically found growing in water colder than their optimal temperature for growth. Modeling studies by Moisan et al. (2002) point out that increased variability in temperature—either periodic or random—will enhance this phenomenon such that as the level of temperature variability increases phytoplankton from warmer regions can dominate. Temperature increases can be detrimental to zooplankton populations by increasing their basil metabolic loss and thereby increasing the level of food required to sustain growth and population success. Many zooplankton species undergo ontogenetic migrations, but little is known as to the environmental keys that turn on or off the diapause resting periods or how environmental variability can or will alter the probability of a successful "deep sleep". Temperature variability can also cause developmental delays or increases in the pelagic eggs of many marine organisms (Capella et al., 1992). These organisms are at increased risk to large decadal variation in temperature if their reproductive habit or behavior has been modified or optimized to known circulation patterns and seasonal succession.

5.3 Horizontal current effects on ecosystem

Horizontal advection is an important determinant for ecosystem variability, especially in the western boundary current region and the KOE region where high-speed currents transport temperature anomalies, change the timing and extent of surface water stabilization, and determine nutrient availability within a mixed layer. Advective effects from incoming Rossby waves on decadal

timescales (Schneider and Miller, 2001; Qiu, 2003) result in temperature anomalies in the KOE SST and variations in upwelling and vertical mixing due to changes in water column stability. In the California Current, long-term observations of hydrography and biological variables yield information on the effects of geostrophic currents on property transports, although interannual timescales dominate the signals (Bograd et al., 2001).

It is noteworthy that advection could be responsible for variations in regional ecosystem structure not only by creating an anomalous physical and chemical environment but also by transporting the organisms themselves. This further complicates climate-ecosystem interactions. An example is the decadal scale population dynamics of the Japanese sardine, *Sardinops melanostictus*, in the Kuroshio Extension (KE) region, a stock that dramatically increased after the 1976–77 shift, corresponding to a cold SST anomaly regime and strong KE current (e.g., Yasuda et al., 1999). The population collapse in the 1990s, when the SST shifted to warm regime and the KE current weakened, was attributed to high mortality of sardine larvae after 1988 (Watanabe et al., 1995). Noto and Yasuda (2003) demonstrated the process with a Lagrangian model study that showed that sardine eggs spawned off southern Japan are transported by the Kuroshio to the KE region, the nursery ground of the larvae. They can expand their distribution during the cold SST periods. A cold water regime preference of Japanese sardine is enigmatic because a warm regime is favorable for eastern Pacific sardine populations (Chavez et al., 2003). It is speculated that mixed layer deepening caused by advection of cold SST increases surface nutrient availability for plankton and consequently increases food availability for sardine larvae (Noto and Yasuda, 1999; Nakata et al., 2001).

Horizontal advection associated with changes in location and intensification of the Alaskan Gyre (Lagerloef, 1995) also is significant in determining decadal scale variation of local ecosystems. Unlike the western boundary region, the respone to the Aleutian Low forcing appears in the surface layer physical environment with no obvious lag in the eastern North Pacific (Miller and Schneider, 2000). A strong Aleutian Low strengthens the Alaskan Gyre, allowing advection of warm southern water during the mid 1970s to 1988. This warm water advection contributed to formation of shallow, stable surface mixed layer and enhanced biological productivity in the Gulf of Alaska (Polovina et al., 1995). Zooplankton biomass doubled (Brodeur and Ware, 1992; Brodeur et al., 1996) and ground fish and salmon catch markedly increased during the 1980s (Hare and Mantua, 2000). However, the link between lower and higher trophic levels is unclear as observed variations are results of complex interaction of various physical and ecological factors (e.g., food availability, predation). Tadokoro (2001)
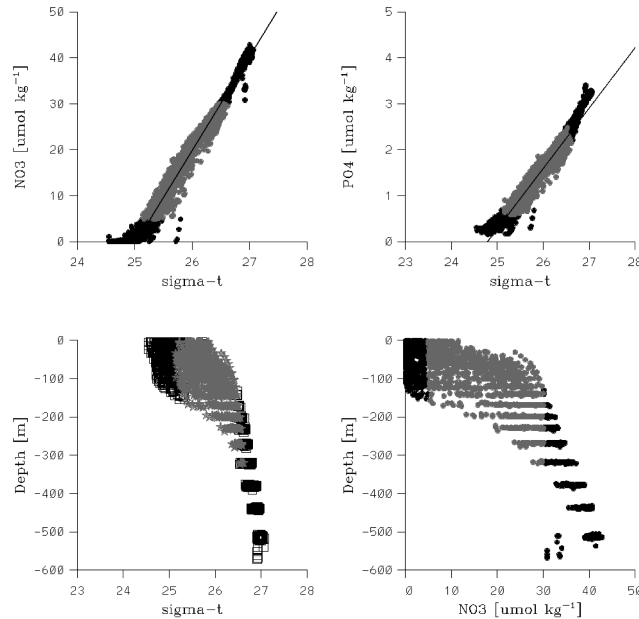
Fig. 6. CalCOFI observations of the relationship between nutrient concentrations and thermocline depth. Grey stars indicate waters having nitrate values between 5 to 30 mmol N m^{-3}, values which are normally in the thermocline/nutricline region. Black boxes indicate other values. (Lower left) Density vs. depth for the year 1999. (Lower right) Nitrate vs. depth for 1999. (Upper left) Nitrate vs. density for 1999. Black line through the grey values is a fit using a PCA method for this is bivariate data. This fit is calculated for this relationship for each of the CalCOFI years with enough good nitrate data and shown in Fig. 4. (Upper right) Phosphate vs. density for 1999, with green line fit as before. These two upper plots show that density and these nutrients are very linearly related in the thermocline.

speculated that northward warm water advection might help to advect mesoscale eddies originating along the continental margin into the open ocean, thereby supplying river flow iron into the Alaskan Gyre. Experimental studies reveal that iron enrichment could increase phytoplankton abundance in the Gulf of Alaska where HNLC conditions usually prevail (Harrison, 2002).

Variations in horizontal currents lead to frontal features that have been shown to be areas of increased primary and secondary production (Smith *et al.*, 1986; Franks and Walstad, 1997). Additionally, mesoscale ocean eddies are argued to enhance the level of nutrient flux into the euphotic zone (McGillicuddy *et al.*, 1998). In either situation, any changes in the eddy energy of the system will modify the frequency that these features exist and therefore can alter the net production of the ecosystem. This can be especially true for pelagic fish populations that tend to forage within these features.

5.4 Vertical mixing and upwelling effects on ecosystem

Upwelling of nutrients is directly influenced by climate variations through changes in longshore wind stress in coastal regions and Ekman pumping in the open ocean regions as well as through changes in vertical mixing. Long-term and large-scale changes in the depth of the thermocline and nutricline can suppress local upwelling. Lagged effects of climate forcing result from the arrival of internal Rossby waves, Kelvin waves or continental shelf waves. These act to complicate the interpretation of the local response, but they also add a predictable component to the system if the dynamics are well understood.

The seasonal cycle of the mixed layer is responsible for mixing deep nutrient-replenished water into the surface layer. Variations in the MLD can increase or reduce the level of new nutrients brought into the euphotic zone and made available for growth. Sensitivity studies on a one-dimensional coupled mixed layer/ecosystem model forced with observational data from OWS Papa (McClain *et al.*, 1996) show that in a 20-year interannual simulation (1960–1980) the variability in annual primary production varied less than 5%. No decadal trends were observed in either the forcing fields (heat fluxes, cloud cover, winds) or the biological variables (phytoplankton, nutrients, zooplankton, etc.). However, it should be noted that OWS Papa is located near the node in the canonical SST pattern and the spatial pattern of the PDO (Fig. 1), thus reducing or eliminating any decadal variability in the forcing fields.

Polovina *et al.* (1995), using the NODC temperature profiles to estimate MLDs, noted that between the peri-

ods of 1960–1976 and 1977–1988 the MLDs in the North Pacific subtropics and the transitions zones deepened by 30–80%, while in the subarctic regions the MLD became shallower by 20–30%. These changes in MLD were consistent with those found by Miller *et al.* (1994a) using results from a basin scale ocean model. With a simple plankton dynamics model, Polovina *et al.* (1995) go on to demonstrate that deeper subtropical and shallower subarctic mixed layers would increase ecosystem production by about 50%. The support from these conclusions is obtained from the widely held belief that the primary production in subtropical regions is nutrient limited while in subarctic regions it is light limited. It is also interesting to note that much of the observed variability in higher trophic levels (fish) is from observations within the subarctic pacific (Francis *et al.*, 1998).

There are several cruxes to the arguments concerning MLD controlling the observed variability in production in the subarctic Pacific. The first is that the spinning up of the subarctic gyre should also increase the level of nutrients available to the phytoplankton while also reducing the predation pressure from zooplankton due to the gyre-scale increase in divergence. The subarctic region is a HNLC region that is thought to be top-down controlled by zooplankton grazing (Frost, 1987).

Other arguments have been made that this HNLC region is limited by iron availability (Martin and Fitzwater, 1988). Additional arguments have been made that while iron might be limiting its specific effect on the diatom species composition may also have a significant effect on the overall plankton ecology in this system (Miller *et al.*, 1991). With respect to decadal variability, any large-scale perturbations in the wind patterns would change the level of iron brought in by atmospheric deposition (Donaghay *et al.*, 1991). This issue was recently investigated by Haigh *et al.* (2001) who noted that while primary production and zooplankton biomass were sensitive to increases in iron availability the net increase in phytoplankton standing stock was minimal. This demonstrates some of the limitations that might be encountered when trying to monitor ecosystem change from satellite data sets. One final note is that the observed decadal variability in open ocean phosphate availability (Karl *et al.*, 2001) may be modulated by wind deposition of iron which is a key trace metal required for support of nitrogen fixation processes (Michaels *et al.*, 2001).

The sensitivity of pelagic ecosystems to decadal variations in MLDs is tightly coupled to variations in thermocline depths that often covary with nutriclines. This is especially true for the CalCOFI region where density and nitrate concentrations vary linearly (Fig. 6). For the divergence regions, such as the subarctic gyre, increased divergence (spinning up of the gyre) brings the thermocline closer to the surface mixed layer. This al-
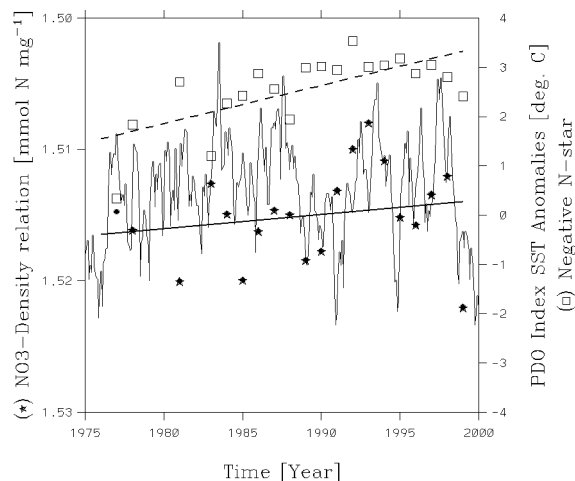

Fig. 7. Long-term changes in the linear relationship between density and nutrients in the thermocline in the CalCOFI region. Time series for the slope of nitrate-density relationship is shown as stars (axis at left). Time series of negative value of N-star (Gruber and Sarmiento, 1997) is shown as squares. Linear fits (univariate) are also drawn to show the long-term trends. Time series for the PDO index is the thin black line (axis to the right).

lows winter deepening of the mixed layer to entrain more nutrients from within the thermocline and import these additional nutrients into the surface layer. A similar effect is expected in subtropical gyres when the gyre spins down. During periods when the thermocline is deep a reverse effect occurs and less nutrients are available for supporting the ecosystem production (Fig. 7). The largest variability in the nitrate-density slope is linked to the PDO index, with a long term but smaller magnitude trend superposed. The quasi-conservative tracer—N* (defined as a linear combination of nitrate and phosphate by Gruber and Sarmiento, 1997) does not seem to hold any PDO signals. The trend in increasing negativity, meaning that phosphate is becoming scarce relative to nitrate, indicates a temporal increase of the amount of denitrification relative to nitrogen fixation going on near the shelves. This should be supported by the oxygen variability that has been observed near the shelves.

A list of the potential changes expected from the various scenarios is given in Table 1. The expectations assume that the changes are short-term in relation to the larger scale basin turnover and nutricline refreshment. How these changes in thermocline depth would further modify the ecosystem depends on if the system was limited by the flux of nutrients from across the nutricline or limited by other processes such as grazing by higher trophic levels or by atmospherically deposited trace metals (iron).

Table 1. Estimates of the expected changes in ocean productivity from a variety of possible decadal changes to MLDs and thermocline depths assuming that production is limited primarily by input of nutrients diffusing through the thermocline.

	Shallower thermocline	Unchanged thermocline	Deeper thermocline
Shallower MLDs	Variable changes to productivity	Decreased productivity	Decreased productivity
Unchanged MLDs	Increased productivity	Unchanged productivity	Decreased productivity
Deeper MLDs	Increased productivity	Variable changes to productivity	Decreased productivity

5.5 Freshwater flux effects on ecosystems

Variations in coastal freshwater flux can be important to ocean climate in several ways. Freshwater flux can modify the stability of the coastal ocean and alter the effectiveness of alongshore upwelling processes. Freshwater fluxes also carry with it high levels of macro- and micronutrients (iron) that may act to modify the total productivity or species composition of the plankton. Variations in freshwater fluxes have been demonstrated to be linked to ENSO events (Cayan and Webb, 1992) and therefore should also have an associated signal with decadal events that influence precipitation events. In addition to effects on nutrient fluxes, changes in freshwater fluxes can add stress to organisms and allow for proliferation or decreases in various diseases related to ecosystem organisms (Hofmann *et al.*, 1995).

5.6 Sea ice effects on ecosystems

Sea ice affects biological productivity in high latitude regions, particularly as a determinant of the timing and extent of the phytoplankton spring bloom (Spindler and Dieckmann, 1994). Fresh water supplied by the sea ice melt in spring forms a stable, thin surface layer around the ice edge in which extensive phytoplankton blooms can occur with available nutrients. The ice edge bloom propagates as sea ice retreats and nourishes the local benthic community and other higher trophic levels. Interannual variations of sea ice coverage is reported to induce a shift in the keystone organisms and the capacity of the biological carbon pump in the Antarctic Ocean (Loeb *et al.*, 1997).

In the North Pacific, decadal scale variations of sea ice extent have been reported in the Bering Sea and Okhotsk Sea. Bering Sea sea ice extent diminished during 1977 to 1988 (Trenberth and Hurrell, 1995), the warm (positive) phase of the PDO index. In the Bering Sea, ice pack formed in polynyas advected southward across the shelf (Macklin, 2001). Therefore, reduction of sea ice area implies decline of spring phytoplankton production for the period in the southeast Bering Sea, while observation results showed increased summertime Chl-a level (Sugimoto and Tadokoro 1997). Spring bloom occurs there in mid-March with sea ice presence, while it will not occur until when water column becomes stratified in May without sea ice presence (Macklin, 2001). The reported summertime Chl a increase could be attributed to a delay in the timing of the spring bloom due to sea ice absence in the large area of the Bering Sea.

Okhotsk Sea sea ice area increased after 1976–77 (Sato, 1992; Iwao and Taguchi, 1999; Sekine, 1999) and both area and duration of coverage markedly declined after 1988 (Aota and Ishikawa, 1993) responding respectively to intensification and attenuation of the Aleutian Low. However, its link to ecosystem variability is not clear because of lack of biological time-series in this region. Okhotsk Sea sea ice dynamics could influence stability of the surface mixed layer in the Oyashio Water by advective supply of cold, less saline ice melt water in spring (Ohtani, 1989). Yet, observations show that the Oyashio mixed layer rather deepened due to strong wintertime wind stress, suppressing spring phytoplankton production for 1976–1988 (Tadokoro, 2001). Thus, sea ice quantity in the Okhotsk Sea is unlikely to influence the lower trophic level ecosystem in the Oyashio Water at least on decadal timescales.

5.7 Intrinsic ecosystem variations

Variability in ecosystem function can be driven by a number of factors, which are modulated by the dominant control within the system. In ecosystems that are bottom-up controlled, one where production by primary producers controls the level of cycling within the system, environmental changes are thought to play a direct role in ecosystem variability. Under such systems, any environmental variability that alters food availability, retention/dispersion, or enrichment processes are thought to directly influence the ecosystem's function and rate of production. Large-scale changes in the northern Pacific fish populations have been well correlated with variability in the inter-decadal climate variability (Mantua *et al.*, 1997) and are an excellent example of such a bottom-up controlled system. As the environment changes it is possible that ecosystems can fluctuate between a number of stable states, as has been suggested from the previous 1976–77 climate shift in the North Pacific (McGowan *et al.*, 2003). In top-down controlled systems, where control is through

predation pressure, environmental changes are thought to play less of a role in modulating the ecosystem's variability. A third form of control is termed "wasp-waist" control, where control occurs through an intermediate but dominants species (Cury *et al.*, 2003). This is thought to be the situation for the upwelling regions off the CCS, where sardine and anchovy populations are observed to dominate under alternating periods at decadal frequencies. Ecosystems are capable of undergoing long-term periodicities through the intrinsic dynamics of the various organisms within it. However, the underlying mechanisms for this variability is thought to be linked to larger scale climate fluctuation because of the large geographic scales of the variability (Lluch-Belda *et al.*, 1992).

5.8 Forcing of climate by attenuation due to phytoplankton

Solar energy in the upper ocean decays exponentially with depth following the Beer-Lambert relation. This attenuation, quantified with the diffuse attenuation coefficient spectrum, is a function of wavelength and determined by the attenuation coefficients of pure seawater and its constituents. In this section we will discuss one type of constituent, the phytoplankton. Since the attenuation of optically pure water is considered constant, changes in chlorophyll biomass are primarily responsible for variations in solar attenuation in open ocean waters. In addition, chlorophyll traps solar energy, in the form of heat, by absorbing solar energy. As the vertical distribution of phytoplankton biomass, with its associated chlorophyll, varies, the vertical distribution of heating will also vary.

Effects on the ocean caused by phytoplankton related heating include variations in the sensible and latent heat exchange with the environment changes in SST, and variations in circulation patterns in both the atmosphere and ocean. Norris and Leovy (1994) found that absorption of solar energy by phytoplankton amplified positive feedback between boundary layer cloudiness and SST; increased SST led to decreases in cloudiness and an increase in insolation ultimately resulting in further SST increases (Sathyendranath *et al.*, 1991). Decreases in solar penetration within the water column also lead to increases in SST and, by trapping the heat near the surface, provide an increased influence in the seasonal and annual temperature cycles of the lowest atmospheric layer.

Recent studies show that the tropical circulation is sensitive to absorption effects of phytoplankton blooms. Murtugudde *et al.* (2002) show that phytoplankton growth in the cold tongue region of the tropical Pacific can warm the mean state by 1°C in an ocean model. Nakamoto *et al.* (2001) showed that phytoplankton blooms alter modeled mixed layer depth and advection in complicated ways that result in anomalous surface temperature fields.

Using a simple model, Timmermann and Jin (2002b) showed that the magnitude of La Niña events is reduced by phytoplankton blooms, thereby explaining a fundamental asymmetry in the ENSO phenomenon. Gildor *et al.* (2003) showed that model intraseasonal oscillations in the tropical Pacific are sensitive to changes in SST induced by the effects of phytoplankton concentration on absorption. Long-term are needed in the tropical Pacific to determine whether these scenarios can occur in nature.

Frouin and Iacobellis (2002) estimated that phytoplankton serve to warm the global atmosphere by up to 0.25°C. Shell *et al.* (2003) forced an atmospheric GCM with an SST pattern that arises from the absorption effect of phytoplankton (Nakamoto *et al.*, 2001) and showed that the amplitude of the global surface-layer atmospheric temperature seasonal cycle increases by roughly 0.5°C. These results support the idea that phytoplankton exert a significant influence on large-scale climate variability.

5.9 Forcing of climate by DMS due to phytoplankton

Blooms of some types of oceanic phytoplankton, especially coccolithophores and small flagellates, produce DMS fluxes to the atmosphere, which serve as cloud condensation nuclei (CCN) in the open ocean (Erickson *et al.*, 1990). Charlson *et al.* (1987) suggested that this ecosystem-climate coupling could moderate global climate and a great deal of research has been executed to determine the magnitude of the effect (Gabric *et al.*, 1998). While the importance of the effect on global climate is still in doubt, its regional impact on atmospheric sensitivity to SST anomalies is virtually unexplored, especially in the context of decadal variability (Miller *et al.*, 2003).

The hot spots of atmospheric sensitivity to SST anomalies in the Pacific are the equatorial Pacific and the KOE regions. Both these regions are capable of producing significant phytoplankton variations on decadal timescales. Hence, the resulting DMS fluxes may alter the cloud formation processes, the storm activity, and the deep convection of the overlying atmosphere. The concomitant teleconnection patterns from the tropical Pacific to midlatitudes and from the KOE region to the central and eastern North Pacific may be fundamentally altered (Miller *et al.*, 2003). Direct observations and numerical experiments are needed to determine if these effects can significantly alter the climate modes of decadal variability in the Pacific sector.

6. Summary and Final Comments

Decadal climate variability in the Pacific Ocean organizes the patterns of biological response in both simple and subtle ways. The dominant patterns of SST variability (the canonical pattern and the KOE pattern; Fig. 1(a) and (b)) affect the mixed-layer structure of the basin

and modulate the fluxes of deep ocean nutrients to the photic zones. The dominant thermocline pattern contributes additional structure to the processes that affect surface layer productivity, including affecting upwelling patterns and upper-ocean stability across the basin. The observed patterns of biological variability in key regions of the Pacific were discussed and explained as much as possible in terms of these patterns. There are many other patterns of decadal variability that have been identified (e.g., Minobe, 2000; Chao *et al.*, 2000; Barlow *et al.*, 2001; Luo and Yamagata, 2002; Lluch-Cota *et al.*, 2003; Wu *et al.*, 2003; Auad, 2003). These also need to be explained physically and linked to biological changes.

Predictable components to the oceanic climate system may be exploited for anticipating ecosystem changes. For example, as wind stress curl drives Rossby waves in the central North Pacific, the waves propagate to the KOE region, arriving some years later and affecting SST and upwelling fields around Japan (Schneider and Miller, 2001). Research needs to be done to determine the full three-dimensional structure of the predictable Rossby waves and how they affect the ecosystem response in that region. Predictable components due to horizontal advection of anomalous conditions may also occur, especially in regions of strong currents. For instance, Freeland *et al.* (2003) identified highly anomalous water masses that subsequently were advected southward along the U.S. west coast (Strub and James, 2003). This type of influence on the ecosystem may be predictable.

On decadal timescales, it is important to identify what portions of the physical climate variations are stochastically driven and what part are oscillatory (Pierce, 2001). Unraveling the processes responsible for decadal variability and how they drive the organization of ocean ecosystems and biogeochemical cycling requires a concerted and organized long-term observational and modeling effort that has only just begun.

Acknowledgements

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