

Chapter 2

Global Change and Oceanic Primary Productivity: Effects of Ocean–Atmosphere–Biological Feedbacks

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Abstract

Our current understanding of how climate change due to increasing greenhouse gases is expected to affect oceanic biology and of how the physical–biological feedbacks may influence the evolving physical climate system is summarized. The primary effects of ocean biology on physical climate include its influence on the carbon cycle, the influence of oceanic phytoplankton on upper-ocean absorption, and the influence of DMS production by phytoplankton on atmospheric aerosols. The primary influences of physical climate on the ocean biology are the influence of aeolian dust deposition and the multitude of ways that community structure can be altered. The focus is on the tropical and midlatitude Pacific Ocean, with results from other ocean basins also noted.

Keywords: global change; climate feedbacks; ocean ecosystem; Pacific Ocean; physical–biological interactions

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1 Introduction

As greenhouse gases continue to increase in the global atmosphere, the coupled climate system will continue to adjust to these changes in radiative forcing in many ways. Among these coupled adjustment processes are possible changes in the physical–biological coupling of the ocean–atmosphere climate and the oceanic ecosystem. In what ways can ocean biology affect changes in physical climate? How might these coupling processes affect the future climate state?

We aim here to summarize the feedback processes among the ocean, the atmosphere, and oceanic biology in the context of how these feedbacks will affect or be altered by global warming scenarios in the Pacific sector. We draw on modeling studies and available observations in key regions of the world ocean, including the tropical Pacific and the midlatitude regions of the northern and southern hemispheres.

2 Ocean Biotic Feedbacks with Centennial Climate Change

Our ability to predict the impacts of global warming is limited by a number of key uncertainties, significant among which is the role of biotic feedbacks (IPCC, 2001). The response of biota in the surface ocean is particularly pertinent and still not well understood. However, the potential for multiple feedbacks between climate, ocean circulation and mixing, and photosynthetic primary production has been manifestly evident for some time (Falkowski et al., 2000; Gildor and Follows, 2002). Indeed, the oceans are estimated to have taken up approximately 30% (with great uncertainty) of CO₂ emissions arising from fossil-fuel use and tropical deforestation between 1980 and 1989, thereby slowing down the rate of greenhouse global warming (Ittekkot et al., 1996).

Although the ocean biota compartment is estimated to contain only around 3 GtC, the flux from the dissolved inorganic reservoir to the particulate organic phase (carbon uptake through primary production) is around 10 GtC yr⁻¹ (Siegenthaler and Sarmiento, 1993). Thus the size of marine biota carbon reservoir is much smaller than the fluxes in and out the reservoir. Elsewhere in the global carbon cycle, the reservoirs are much larger than the fluxes. This implies that any changes in the activity of this reservoir can mean substantial changes in the fluxes to related reservoirs. An especially important flux in the oceans is the burial of particulate organic carbon in marine sediments, which removes atmospheric CO₂ for prolonged time periods. Fig. 1 schematizes the major carbon reservoirs and flux directions in the global carbon cycle.

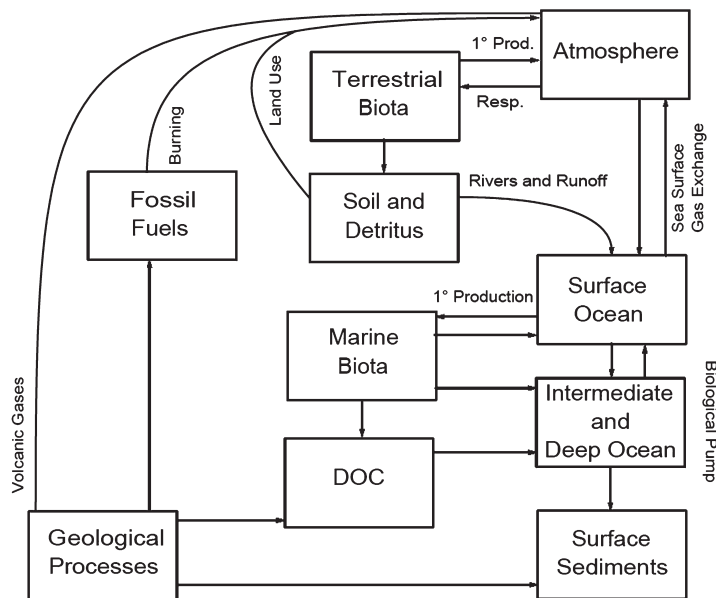


Figure 1: Schematic diagram of the major reservoirs and flux directions of the global carbon cycle.

Photosynthesis, the major process by which marine biota sequester CO_2 , is largely controlled by the availability of macronutrients and trace elements such as iron (de Baar et al., 1995; Behrenfeld et al., 1996; Coale, 1996; Falkowski et al., 1998). Changes in freshwater runoff or increases in aeolian dust transport resulting from climate warming could change the inputs of nutrients and iron to the ocean, thereby affecting CO_2 sequestration.

Climate change can also cause shifts in the structure of biological communities in the upper ocean – for example, between coccoliths and diatoms. In the Ross Sea, diatoms (primarily *Nitzshia subcurvata*) dominate in highly stratified waters, whereas *Phaeocystis antarctica* dominate when waters are more deeply mixed (Arrigo et al., 1999). Changes to ocean stratification could impact species composition and alter the downward fluxes of organic carbon and consequently the efficiency of the biological pump.

Several coupled atmosphere-ocean models have been used to project the effect of climate change on marine biota (Sarmiento et al., 1998; Joos et al., 1999; Gabric et al., 2003; Pierce, 2003). These models include some or all of the processes associated with carbonate chemistry and gas exchange, physical and biological uptake of CO_2 , and changes in temperature, salinity, wind speed, and ice cover. They account for simple changes in biological productivity, but not for changes in external nutrient supply, or changes in the biogeography of planktonic species, which is a major deficiency as they thus cannot simulate more complex biological feedbacks (Gabric et al., 2003).

1 The range of model estimates of the climate change impact is dependent
3 on the choice of scenario for atmospheric CO₂ and on assumptions concern-
5 ing marine biology. At high CO₂ concentrations, marine biology can have a
7 greater impact on atmospheric CO₂ than at low concentrations because the
9 buffering capacity of the ocean is reduced (Sarmiento and Quéré, 1996).
11 Although the impact of changes in marine biology is highly uncertain and
13 many key processes are not included in current models, sensitivity studies
15 can provide approximate upper and lower bounds for the potential impact of
17 marine biology on future ocean CO₂ uptake. A sensitivity study of two ex-
19 treme scenarios for nutrient supply to marine biology gave a range of 8–25%
21 for the reduction of CO₂ uptake by mid-21st century (Sarmiento et al., 1998).
23 This range is comparable to other uncertainties, including those stemming
25 from physical transport.

17 **3 Ocean–Atmosphere–Ecosystem Feedback** 19 **Processes**

21 Climate feedback processes can be broadly divided into two categories: geo-
23 physically mediated, where only the physical climate system is important,
25 and biogeochemically mediated, where various biospheric and geochemical
27 components become important as well (Lashof, 1989). Examples of the first
29 category include water vapor feedback, the sea-ice/solar albedo feedback and
31 various cloud feedbacks. Examples of the second category include changes in
33 upper-ocean absorption due to phytoplankton and changes in community
35 structures that affect CO₂ cycling and storage. The biogeochemical feed-
37 backs that affect changes in temperature, radiation, and moisture are the
39 most important since they have the most direct link to the physical climate
41 system (Woodwell et al., 1998).

43 Currently, most global climate models do not include biogeochemically
mediated feedback processes explicitly, such as the role of ocean ecosystems
and its impact on carbon cycle and climate variability. Important new ad-
vances, however, have allowed new progress in this direction as summarized
recently by Miller et al. (2003). We next explain the key processes by which
ocean biology may influence the physical climate system over timescales
relevant to global warming. Besides the effect of oceanic biology on altering
carbon cycling and sequestration (Boyd and Doney, 2003), these processes
include the effects of phytoplankton on upper-ocean absorption of radiation
and the production of DMS aerosols by certain types of phytoplankton.
Physical forcing also affects the ecosystem in important ways. These include
changing the deposition of aeolian dust, which affects the limiting nutrient
iron in the ocean and changing community composition by altering the en-

1 vironment in the ocean. The next sections will summarize these issues,
3 including a final section on the local response of the North Pacific Ocean.

5 **4 Absorption of Radiation by Phytoplankton in** 7 **the Upper Ocean**

9 The effect of the absorption of solar energy by phytoplankton on upper-ocean
11 thermal properties has been the subject of research for the past 20 years.
13 While absorption of solar energy is dominated by absorption from seawater
15 itself in many open ocean regions, the variability in the absorption and
17 distribution of solar energy into the upper layers of the open ocean is con-
trolled primarily by phytoplankton pigment concentrations (Platt, 1969;
Smith and Baker, 1978). Lewis et al. (1983) were the first to demonstrate
that nonuniform vertical distributions of phytoplankton pigments cause
variations in local heating, and, under certain vertical chlorophyll profile
conditions, could support the development of a thermal instability within
the water column.

19 Initial attempts (Paulson and Simpson, 1977) at addressing the effects of
21 varying water quality types [as described by Jerlov (1968)] on the attenu-
23 ation of irradiance in the ocean lead to a simple parameterization that
25 characterizes absorption between the longwave and shortwave (visible)
27 bands of solar energy using different e-folding scales, and set the e-folding
29 scale of the shortwave band dependent on the water quality type. This pa-
31 rameterization has commonly been used to provide buoyancy forcing in one-
dimensional, ocean surface mixed-layer models (Price et al., 1986; Schudlich
and Price, 1992). More sophisticated methods to estimate the solar energy
flux into the ocean resolve the depth- and wavelength-dependent spectral
diffuse attenuation coefficients (Siegel and Dickey, 1987; Morel and Antoine,
1994), and Siegel and Dickey (1987) have shown that this method greatly
improves the ability to compare observed irradiance fields to model esti-
mates.

33 The first work to directly address the link between ocean thermody-
35 namics and bio-optical processes (Simonot et al., 1988) coupled the bulk
37 mixed-layer model of Gaspar (1985, 1988) to a simple, nonspectral, diffuse
39 attenuation model for solar energy attenuation and a six-component eco-
system model (Agoumi et al., 1985). Results on simulations of the seasonal
cycles at Ocean Weather Station Romeo show that the phytoplankton sea-
sonal cycle has a significant impact on sea surface temperature evolution.

41 While early modeling studies all agreed that chlorophyll attenuation
43 plays an important role in ocean physics, few direct observations had been
available to confirm this. However, during the coastal transition zone field
study along the California coast, Ramp et al. (1991) interpreted observations

1 of a surface warming feature during a wind relaxation event to be caused by
2 patchiness in the near surface chlorophyll distribution. The first notion that
3 this biological–physical process acted on equatorial ocean regions was pre-
4 sented by Sathyendranath et al. (1991), who argued that chlorophyll
5 patches were important driving mechanisms for variations in sea surface
6 temperature. Such sea surface temperatures (SST) anomalies have been
7 shown (Kershaw, 1985) to influence the evolution of Arabian Sea monsoons.
8 Additionally, Kahru et al. (1993) presented evidence from AVHRR satellite
9 analysis and *in situ* observation to show that cyanobacteria blooms in the
10 Baltic Sea were responsible for elevating the SST to 1.5°C. Global analysis of
11 the ocean color fields in the tropical Pacific Ocean (McClain et al., 2002) also
12 verified that enhanced chlorophyll regions were linked with enhanced sur-
13 face-layer heating. Further evidence of the impact of phytoplankton on the
14 evolution of mixed layers is presented by Stramska and Dickey (1993), who
15 used bio-optical observations from a mooring off Iceland in conjunction with
16 a version of the Mellor-Yamada two-and-a-half layer mixed layer model
17 (Mellor and Yamada, 1982) to show that the importance of this coupling is
18 most significant in regions of high chlorophyll and weak vertical mixing.
19 One such region is the equatorial region of the ocean, where high solar
20 fluxes are collocated with low wind speed “doldrums” (Fig. 2a) and high
21 chlorophyll equatorial upwelling regions (Fig. 2b).

22 In the western warm pool (WWP) region of the Pacific Ocean, Siegel et
23 al. (1995) demonstrated that the amount of solar radiation penetrating
24 through the bottom of the mixed layer ($\sim 23 \text{ W m}^{-2}$ at 30 m) is a large frac-
25 tion of the net air–sea heat flux ($\sim 40 \text{ W m}^{-2}$). Following a period of sustained
26 westerly wind burst and a corresponding near 300% increase in mixed-layer
27 chlorophyll concentrations, the resulting biologically mediated increase in
28 solar energy attenuation created a decrease in energy flux across the mixed
29 layer (5.6 W m^{-2} at 30 m) and supported a mixed-layer heating rate of
30 $0.13 \text{ }^\circ\text{C}$ per month. In the same year, Ramanathan et al. (1995) believed that
31 a discrepancy occurred within the computed heat balance of the ocean-at-
32 mosphere energy budget in the western equatorial Pacific and that this
33 discrepancy was due to “A Missing Physics” which would modify the manner
34 and importance of cloud absorption of solar energy. Arguments were pre-
35 sented that this Missing Physics was in fact related to the manner in which
36 solar radiation penetrates through the bottom of the mixed layer in this
37 clear water region (M. R. Lewis, private communication). Further evidence
38 has largely dismissed the claims of Ramanathan et al. (1995), and the im-
39 portance of properly attenuating solar energy into the water column is now
40 widely accepted.

41 The importance of characterizing the penetrative fluxes of solar energy
42 through the upper-ocean mixed layer and into the permanent pycnocline
43 prompted Ohlmann et al. (1996) to carry out a global analysis of the mag-

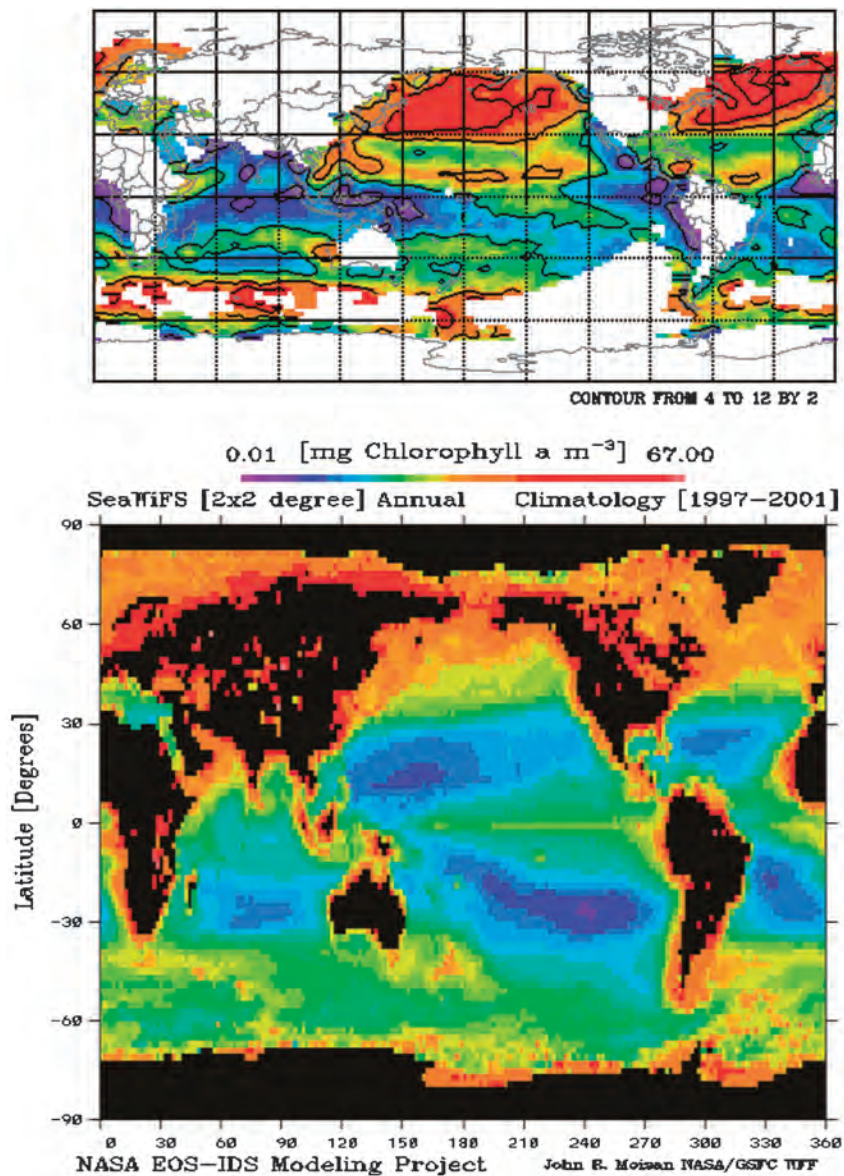
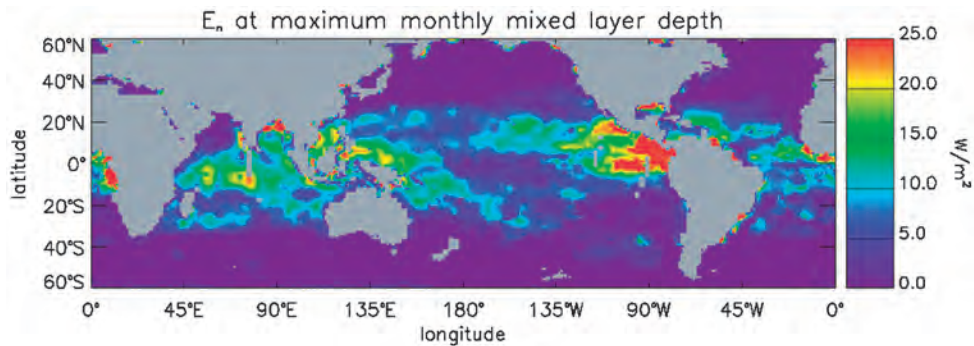


Figure 2: (a) January mean of Oberhuber atlas surface wind field. Note the low wind speeds near the equatorial Pacific and Indian Ocean regions. (b) SeaWiFS annual mean 2×2 degree binned climatology. Note the high chlorophyll values in the eastern equatorial Pacific.

41 nitude of this flux. The global map of these fluxes (Fig. 3) shows high net
43 net solar fluxes ($10\text{--}25 \text{ W m}^{-2}$) in the equatorial Pacific and Indian Ocean re-
gions. The values are highest at the eastern regions of the equatorial Pacific



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Figure 3: Modeled climatological values of the net solar flux at the base of the deepest monthly mixed layer (W m^{-2}). Values correspond to solar fluxes entering the permanent pycnocline. Largest values exist where the deepest monthly mixed layer and chlorophyll concentration are low and solar flux is high. From Ohlmann et al. (1996).

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Ocean, where heat below the pycnocline is transported west to the Pacific arm pool regions. How variations in these fluxes are linked to El Niño southern oscillations (ENSO) dynamics is still unknown, but storage of heat below the mixed layer can tie up heat energy until winter ventilation/mixing processes entrain it back into the mixed layer. Further demonstrations of the importance of bio-optical forcing (Ohlmann et al., 1998), using data collected from the western Pacific warm pool during TOGA-COARE and mixed-layer model simulations, noted also that increases in the penetrative heat loss to below the mixed layer resulted in a destabilization of the thermocline and a deepening of the mixed layer – creating a feedback mechanism for ocean heat flux and mixed-layer depths (MLDs) that are modified through chlorophyll concentrations.

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An additional link was made between clouds and ocean heat flux processes by Siegel et al. (1999) who show that under cloudy sky conditions the near UV to green fraction of the solar spectrum is less absorbed than the rest of the solar energy spectrum. This allows a greater fraction of the total energy to penetrate further into the water column. At 0.1 m depth, this relative increase can be as high as a factor of 2 and likely influences the diurnal heat balance by altering the upper most ocean layer daily heat balances, and could alter the local heat budgets on longer timescales when taking into account the effects on the ocean-atmosphere latent, sensible and back radiation terms. A more recent effort using observations from the hyperspectral ocean dynamics experiment (HYCODE) and a radiative transfer model shows that the rate of heating in a coastal region water column can increase by $\sim 0.2^\circ\text{C} (13\text{ h})^{-1}$ during high chlorophyll conditions (Chang and Dickey, 2003).

1 The majority of the research in the 1990s focused on demonstrating the
2 importance of bio-optics in modifying the vertical flux of heat in the upper
3 ocean. Links between bio-optical forcing of the upper-ocean thermal struc-
4 ture and horizontal momentum forcing only began to appear in the early
5 part of this millennium. Edwards et al. (2001, 2004) used steady-state forms
6 of the momentum equation in conjunction with an analytical description of a
7 high concentration band of phytoplankton biomass (chlorophyll) to ascertain
8 the effect of chlorophyll on ocean circulation patterns. The results demon-
9 strated that the presence of chlorophyll in the water has an impact on ocean
10 circulation, creating both horizontal currents and bands of upwelling and
11 downwelling in regions near the chlorophyll/biomass front. Gildor et al.
12 (2003), in another modeling study, used a simple atmospheric model for
13 climate coupled to a nitrogen–phytoplankton–zooplankton (NPZ) model
14 (Edwards and Brindley, 1999) to demonstrate that intraseasonal variations
15 in SST and precipitation could be forced by inherent oscillations of an ec-
16 osystem.

17 Recent results from coupled circulation/bio-optical models have demon-
18 strated the significance of biological feedbacks with the ocean climate.
19 Phytoplankton pigment concentrations derived from the coastal zone color
20 scanner (CZCS) were used by Nakamoto et al. (2000) to force an isopycnal
21 ocean circulation model coupled to a mixed-layer model to show that the
22 higher chlorophyll concentrations in October versus May increased the
23 amount of solar energy absorption and the rate of heating in the upper
24 ocean. These changes lead to a decrease in MLDs, a decrease in water tem-
25 peratures beneath the mixed layer, and an increase in surface mixed-layer
26 temperature. Comparison simulations of the equatorial Pacific (Nakamoto
27 et al., 2001) using a similar coupled isopycnal-mixed layer ocean circulation
28 model and forced with and without chlorophyll (CZCS-derived pigments)
29 demonstrated that the presence of the chlorophyll leads to shallower mixed
30 layer in the equatorial Pacific, which generates anomalous westward geo-
31 strophic currents north and south of the equator. In the western equatorial
32 Pacific, the anomalous currents enhance the equatorial undercurrent
33 (EUC). The biologically enhanced EUC leads to anomalous upwelling in
34 the eastern equatorial Pacific, while the spatially averaged SST over the
35 Pacific increases due to heat trapped by phytoplankton in the upper ocean.
36 Using sea-viewing wide field-of-view sensor (SeaWiFS)-derived chlorophyll
37 pigment data in the MIT global ocean model, Ueyoshi et al. (2003) confirmed
38 the process described by Nakamoto et al. (2001) whereby chlorophyll mod-
39 ulates oceanic heat uptake by radiation and subsequently generates biolog-
40 ically induced currents in the equatorial Pacific.

41 Shell et al. (2003) forced an atmospheric general circulation models
42 (GCM) with the SST pattern that arises from this phytoplankton effect and
43 showed that the amplitude of the global surface-layer atmospheric temper-

1 ature seasonal cycle increases by roughly 0.5°C. Frouin and Iacobellis (2002)
2 estimated that phytoplankton serves to warm the global atmosphere by up
3 to 0.25 °C, supporting the idea that phytoplankton exerts a significant in-
4 fluence on large-scale climate variability. Oschlies (2004) showed that sur-
5 face heat fluxes act as a negative feedback to reduce the absorptive warming
6 effects of phytoplankton in the upper ocean of a fully interactive physi-
7 cal–biological model. Manizza et al. (2005) used a fully coupled physical–bi-
8 ological ocean model to show that phytoplankton biomass amplifies the
9 seasonal cycles of SST, MLDs, and ice cover by roughly 10%.

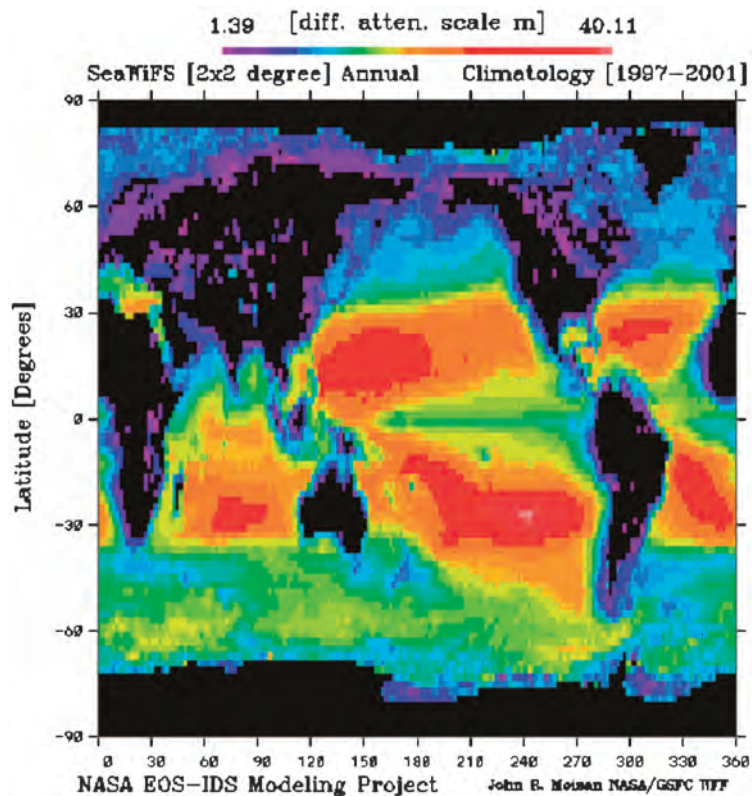
10 Attenuation of solar energy into the ocean using diffuse attenuation
11 coefficients has been used for a variety of ocean modeling studies. Rochford
12 et al. (2001) developed a global field for the diffuse attenuation k_{PAR} of pho-
13 tosynthetically available radiation (PAR) – the visible portion of the solar
14 energy spectrum that is not absorbed in the first several centimeter of water
15 column – using data from the SeaWiFS. The diffuse attenuation field was
16 used in the finite depth version of the NLOM global ocean circulation model
17 with an embedded mixed layer to determine the sensitivity of the model
18 solutions to the diffuse attenuation fields. The results demonstrated that
19 using the derived SST prediction improved in the low latitude regions but
20 the MLD predictions showed no significant improvement. In addition, using
21 a constant clear ocean k_{PAR} value of 0.06 m⁻¹ produces reasonable results for
22 much of the global ocean regions.

23 In a similar study using a primitive equation, global ocean circulation/
24 mixed layer model forced with spatially varying radiation attenuation co-
25 efficients derived from CZCS data, Murtugudde et al. (2003) show that the
26 results from such coupled models can be counterintuitive. For instance, in
27 the eastern equatorial Pacific, where the presence of high chlorophyll leads
28 to strong attenuation of solar energy, realistic solar energy attenuation
29 leads to increased subsurface loss of solar energy, increased SST, deeper
30 mixed layers, reduced stratification, and horizontal divergence (upwelling/
31 downwelling). Timmermann and Jin (2002), using a dynamic ENSO model,
32 point out that eastern equatorial Pacific ocean chlorophyll blooms during La
33 Nina periods create a temperature regulating negative feedback that redi-
34 stributes heat into the surface layer and the associated results from the
35 air–sea coupling dampens the La Nina conditions. This mechanism is
36 thought to counter the positive Bjerknes atmosphere–ocean feedback that
37 links La Nina events with stronger trade winds that force stronger upwell-
38 ing leading to the intensification of La Nina conditions.

39 More sophisticated attempts to link the role of ocean biological feedback
40 mechanisms are just beginning to emerge and support the notion that bi-
41 ological effects enhance ENSO variability. Marzeion et al. (2005) used a
42 primitive equation ocean model with a dynamic ocean mixed layer and a
43 nine-component ecosystem model coupled to an atmospheric mixed-layer

1 model and a statistical atmospheric model to investigate the feedback be-
 2 tween chlorophyll concentrations and the ocean heat budget in the tropical
 3 Pacific. The results from this study supported the earlier conclusions by
 4 Timmermann and Jin (2002) of a bioclimate feedback mechanism and ear-
 5 lier results describing the possible effects on the surface ocean currents
 6 (Murtugudde et al., 2003). The results present a scenario where subsurface
 7 chlorophyll concentrations force changes in subsurface heating rates and
 8 leading to changes in subsurface heating, mixed-layer deepening, altera-
 9 tions in surface ocean currents, and ultimately supporting an eastern Pacific
 10 surface warming.

11 The most recent version of the community climate system model under
 12 development at NCAR and the NASA MOM4 model is making use of the
 13 observed spatially varying diffuse attenuation obtained from ocean color
 14 estimates (Fig. 4). Future global climate simulations will be taking this
 15 physical–biological feedback mechanism into account (Ohlmann, 2003).



41 Figure 4: The 2×2 degree annual mean SeaWiFS-derived diffuse atten-
 42 uation coefficient [m^{-1}] field for PAR.
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5 Production of Atmospheric DMS by Oceanic Phytoplankton

Dimethylsulfide (DMS) is the most abundant form of volatile sulfur (S) in the ocean and is the main source of biogenic reduced S to the global atmosphere (Andreae and Crutzen, 1997). The sea-to-air flux of S due to DMS is currently estimated to be in the range (15–33) Tg S yr⁻¹, constituting about 40% of the total atmospheric sulfate burden (Chin and Jacob, 1996). At the hemispheric scale (Gondwe et al., 2003) estimate that seawater DMS contributes 43% of the mean annual column burden of non-sea-salt sulfate (nss-SO₄²⁻) in the relatively pristine southern hemisphere, but only 9% in the northern hemisphere, where anthropogenic sulfur sources are overwhelming.

During its synthesis and cycling in the upper ocean, DMS is ventilated to the atmosphere, where it is rapidly oxidized to form nss-SO₄²⁻ and methanesulfonate (MSA) aerosols. Sulfate aerosols (of both biogenic and anthropogenic origin) play an important role in the earth's radiation balance both directly through scattering, absorption, and reflection of solar and terrestrial radiation, and indirectly, by modifying cloud microphysical properties (Charlson et al., 1992). The flux of DMS from the ocean to the atmosphere is an important concern for atmospheric modelers since the net effect of DMS is believed to be a cooling effect for the global climate (Kiene, 1999). While wind forcing is known to control the piston pumping velocity of DMS gas across the air-sea interface (Liss and Slater, 1974), the percent yield of DMS from DMSP ($100 \times \text{DMS}_{\text{Production Rate}} / \text{DMSP}_{\text{Consumption Rate}}$) is well correlated to MLDs (Simó and Pedrós-Alló, 1999). Combining this relationship with the global climatologies of MLDs produces a map demonstrating the seasonal and global variability of this efficiency (Fig. 5). Combining these estimates with estimates of the air-sea flux of DMS (Bates et al., 1987a) may have some potential for gaining additional insight in the level of net DMS production in the ocean. In addition to seasonal variations in DMS production, ecosystem changes in the Pacific equatorial regions are now known to undergo spatial and temporal variations that are linked to larger scale climate variations such as ENSO (Karl et al., 1995). Such variability should also impact the production level of DMS and its associated air-sea flux.

Various phytoplankton species synthesize differing amounts of dimethylsulfoniopropionate (DMSP), the precursor to DMS. The function of DMSP in algal physiology seems to be varied, and it is thought to act as an osmolyte, a cryoprotectant, and also relieve oxidative stress in the algal cell (Kirst et al., 1991; Liss et al., 1993; Stefels, 2000; Sunda et al., 2002). In general, coccolithophorids and small flagellates have higher intracellular concentrations of DMSP.

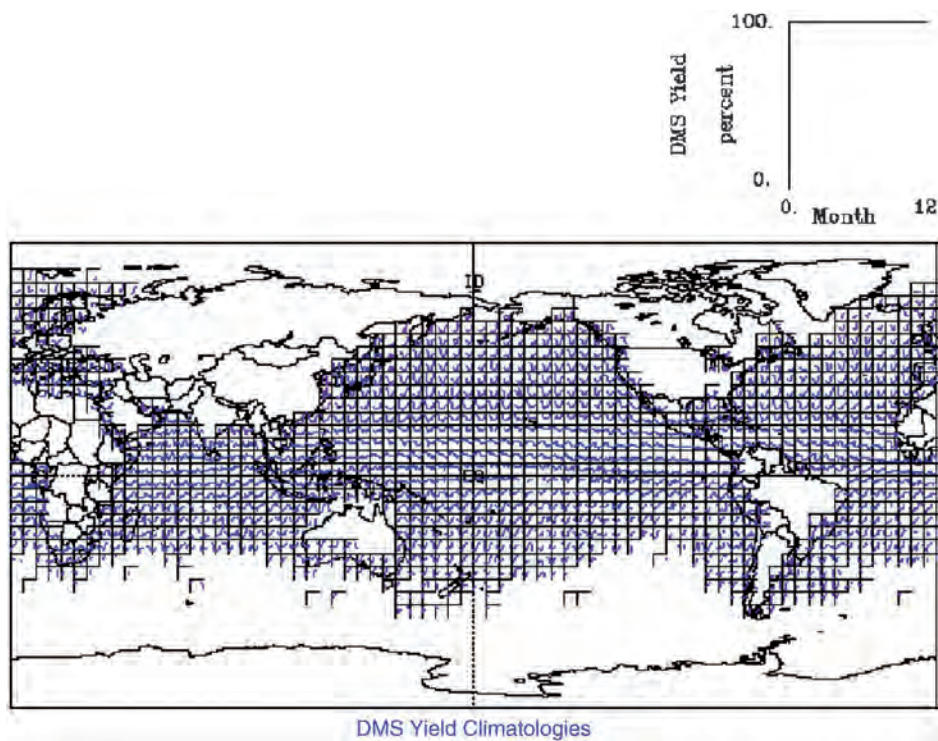


Figure 5: DMS percent yield climatologies estimated using observed MLD climatologies from the NODC XBT data set and the Simo and Pedrós-Alló (1999) DMS yield relationship.

It is well established that the oceans are highly supersaturated in DMS with respect to atmospheric concentrations (Barnard et al., 1982; Liss et al., 1993). In fact, DMS is so supersaturated that recent field experiments on air-sea CO_2 gas flux have used the flux of DMS as a proxy signal to parameterize the physical process of air-sea gas transfer (Dacey, private communication, 2003). Vertical profiles of DMS in the Sargasso Sea (Dacey et al., 1998) show a marked subsurface 10 m maximum. A comparison of the depth-integrated annual cycle of DMS, chlorophyll, and primary production rates in this region shows that DMS concentrations peak in late summer (August) when both chlorophyll and primary production rates are lower than their earlier spring maximum values – suggesting that DMS production by phytoplankton is not directly linked to photosynthetic processes and may be due to release from grazing by zooplankton (Leck et al., 1990). However, correlations between algal biomass and DMS concentrations have been found for dinoflagellate and coccolithophore blooms (Leck et al., 1990). Because of these obvious complexities, attempts to model the production of DMS (Gabric et al., 1993) in ecosystem models have included both direct

(primary production) and indirect (grazing) sources. Refer to Lee et al. (1999) for a recent review of DMS in aquatic environments.

Shaw (1983) and then Charlson et al. (1987) [notably called the “CLAW Hypothesis,” derived from the first letters of the author’s last names], postulated links between DMS, atmospheric sulfate aerosols, and global climate. It was hypothesized that global warming would be accompanied by an increase in primary production, and biogenic production of DMS-derived sulfate aerosols leading to increased scattering, more cloud condensation nuclei (CCN), and brighter clouds. Such changes in the atmosphere’s radiative budget would cool the earth’s surface and thus stabilize climate against perturbations due to greenhouse warming. While phytoplankton is the protagonists in this feedback loop, recent advances in understanding the complex cycle of DMS suggest that it is the entire marine food web (Fig. 6) that determines net DMS production and not just algal taxonomy (Simó, 2001).

The emission of DMS and aerosol particle concentrations is well correlated across varying latitudes and seasons (Bates et al., 1987b). However, Schwartz (1988), in a comparison between southern (SH) and northern hemisphere (NH) cloud albedo records, argues that the CLAW hypothesis is not valid since the anthropogenically introduced sulfur aerosols in the NH should have created a noted increase in cloud albedo over the past century

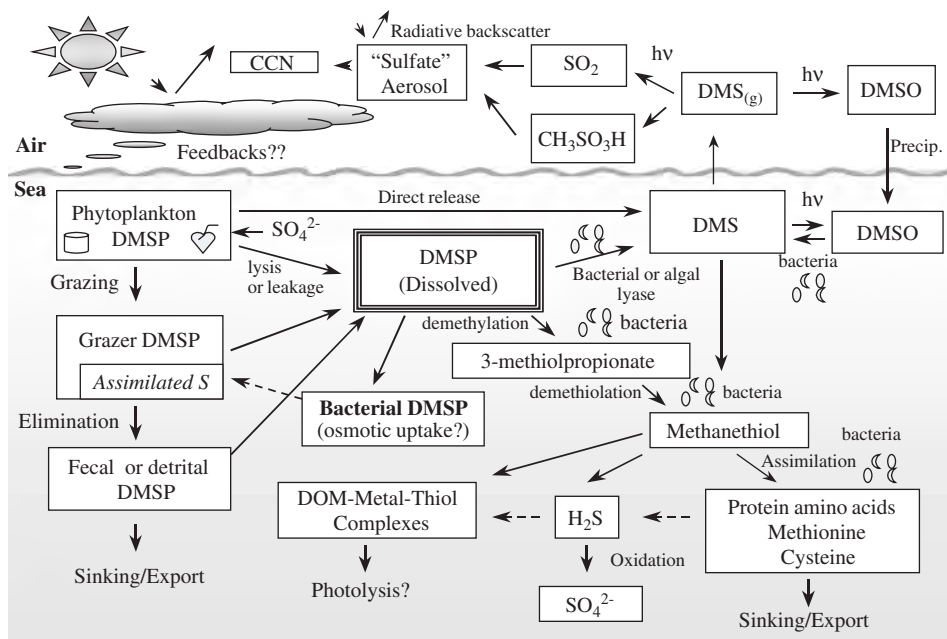


Figure 6: Conceptual model of the cycling of DMSP and DMS in the upper ocean.

1 and none were observed. The debate on the CLAW hypothesis continues to
2 date (Sherwood and Idso, 2003).

3 The DMS-climate feedback hypothesis has stimulated a very significant
4 research effort. Several large-scale studies inspired by the International
5 Global Atmospheric Chemistry program (IGAC) have addressed aspects of
6 the DMS–aerosol–climate connection, including ASTEX/MAGE (Huebert et
7 al., 1996), ACE-1 (Bates et al., 1998), and AOE-91 (Leck et al., 1996). A
8 global database of DMS seawater concentrations and fluxes has been com-
9 piled (Kettle et al., 1999; Kettle and Andreae, 2000), and more recently, a
10 simple empirical algorithm relating DMS seawater concentration to the
11 oceanic MLD and surface chlorophyll concentration has also been derived
(Simó and Dachs, 2002).

12 Notwithstanding this progress, the quantitative evaluation of the DMS-
13 climate hypothesis remains a daunting challenge. This is due in part to the
14 need to integrate knowledge across the traditional disciplinary boundaries
15 of ecology, oceanography, and atmospheric science but also due to our in-
16 complete understanding of the DMS marine production cycle.

17 General circulation models predict the planet’s mean temperature will
18 increase under the “business as usual” scenario (Houghton et al., 1996). The
19 most recent estimate of average warming for a doubling of CO₂ is 3.3 ± 0.8 °K
20 (Grassl, 2000). However, there is strong spatial variation in this perturba-
21 tion, with large temperature and salinity changes predicted to occur in the
22 polar oceans (e.g., Hirst, 1999). The associated warming and salinity re-
23 duction is generally accompanied by a shallowing of the oceanic mixed layer,
24 and stronger illumination of the upper water column, both of which can
25 affect the food-web dynamics and consequently DMS production (Gabric et
26 al., 2001a). It is pertinent to note that the Simó and Dachs’ algorithm em-
27 ploys an inverse relation between MLD and DMS concentration, suggesting
28 that DMS seawater concentration is likely to increase under global warm-
29 ing.

30 Attempts to assess the direction and magnitude of the DMS-climate
31 feedback (Foley et al., 1991; Lawrence, 1993; Gabric et al., 2001b) indicate a
32 small-to-moderate negative feedback on climate (stabilizing), with magni-
33 tude of order 10–30%, and considerable regional variability. The results of
34 the use of GCM data to force a DMS model in the Antarctic Ocean under a
35 global warming scenario suggests that significant perturbation to the DMS
36 flux will occur at high latitudes (Gabric et al., 2003). Fully coupled climate
37 and biogeochemistry models (Joos et al., 1999; Cox et al., 2000) are the next
38 step in further unraveling the DMS-climate link.

41

43

6 Deposition of Aeolian Dust on the Ocean by the Atmosphere

Aeolian dust deposition over the oceans provides a biogeochemical link between climate change and terrestrial and marine ecosystems (Ridgwell, 2002). A major natural source of new iron to open ocean surface waters is continentally derived aeolian dust, which supplies about three times the fluvial input (Duce and Tindale, 1991). *In situ* iron-fertilization experiments have been conducted in both the equatorial Pacific [IronEx I (Martin et al., 1994) and IronEx II (Coale, 1996) and Southern Ocean (SOIREE (Boyd and Law, 2001))]. On all three occasions, raising the iron level in the water by a few nanomoles per liter produced a significant increase in phytoplankton biomass. During IronEx II, the increase was at least an order of magnitude. Iron-limited high nutrient low chlorophyll (HNLC) regions comprise approximately 30% of the world ocean and include the Southern Ocean (de Baar et al., 1995). The majority of iron deposition to the ocean occurs in the NH and is principally associated with dust export from the major arid zones such as the Sahara and Taklamakan Deserts (Fig. 7). The North Atlantic and North Pacific Oceans receive 48% and 22% of global iron deposition to the oceans, while the Indian Ocean (principally in the Arabian Sea) receives 18% and the Mediterranean Sea 4%. The South Atlantic and South Pacific

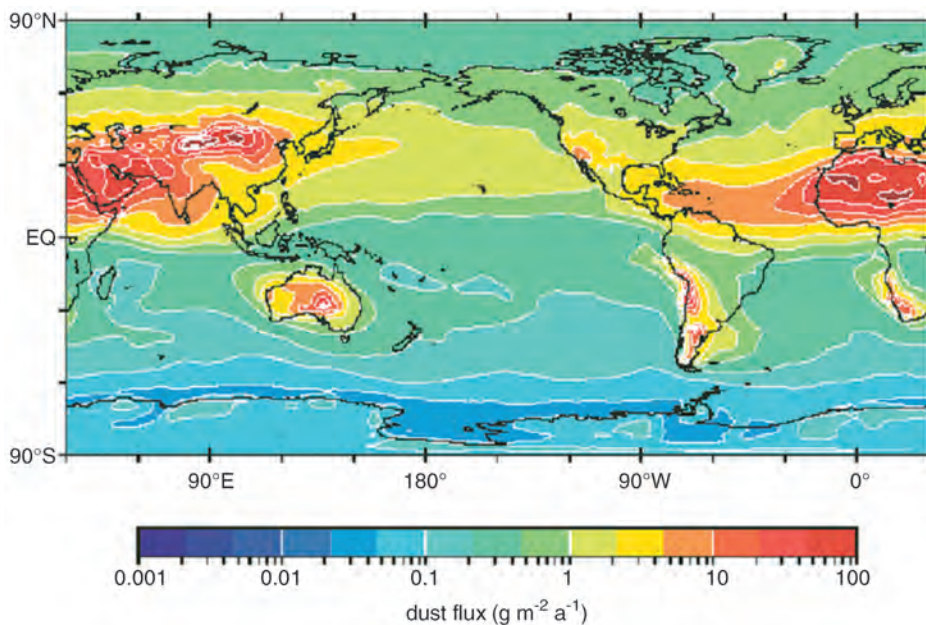


Figure 7: Contemporary annual mean dust deposition rate (Ginoux et al., 2001).

1 Oceans receive only 4% and 2%, respectively, while the polar regions in both
2 hemispheres also receive very low iron inputs, with the Arctic receiving only
3 0.9% and the Antarctic 0.5% (Gao et al., 2001).

4 The subantarctic Southern Ocean is an HNLC region, and it has been
5 suggested that deep mixing and the availability of iron limit primary pro-
6 duction. Australian mineral dust is high in iron content and can be trans-
7 ported over the Australian sector of the subantarctic Southern Ocean,
8 particularly during the austral spring and summer when dust storm fre-
9 quency in southern Australia is maximal. Recently, Gabric et al. (2002)
10 reported evidence for a coupling between satellite-derived (SeaWiFS) aer-
11 osol optical thickness and chlorophyll concentration in the upper ocean. The
12 coupling was evident at monthly, weekly, and daily timescales. The shorter
13 timescale coupling supports the hypothesis that episodic atmospheric de-
14 livery of iron is stimulating phytoplankton growth in this region.

15 Long-term climate variability may also be related to variation in dust
16 deposition rates. Evidence for a possible influence on the glacial–interglacial
17 variability of atmospheric CO₂ comes from the observed changes in dust
18 deposition, recorded in the 420 kyr Vostok Antarctic ice core (Petit et al.,
19 1999). The concentration of dust contained within the ice exhibits a series of
20 striking peaks against a background of relatively low values. Intriguingly,
21 the occurrence of these peaks correlates with periods of particularly low
22 atmospheric CO₂ values. It has been hypothesized that enhanced dust sup-
23 ply to this region during the last glacial could have driven a more vigorous
24 oceanic biological pump with consequent draw-down of atmospheric CO₂
25 (Martin, 1990).

26 Numerical models of the global carbon cycle have since demonstrated
27 that realistic increases in the strength of the biological pump in the South-
28 ern Ocean are unable to explain glacial atmospheric CO₂ mixing ratios as
29 low as ~190 ppm. However, of the total ~90 ppm deglacial rise in atmos-
30 pheric CO₂, the initial 40–50 ppm occurs extremely rapidly (within just
31 ~3 kyr) and up to 10 kyr before the collapse of the NH ice sheets. Predictions
32 of a carbon cycle model that explicitly accounts for the biogeochemical cy-
33 cling of Fe in the ocean, confirm that changes in the aeolian supply of Fe to
34 the Southern Ocean may be at least partly responsible for these particular
35 features of the CO₂ record (Watson et al., 2000).

36 Interestingly, it has also been proposed that aeolian delivery of iron can
37 also influence the oceanic sulfur cycle and the oxidation of DMS in the
38 remote marine atmosphere. Zhuang et al. (1992) report that over 50% of the
39 total iron present in remote marine aerosols is in the soluble Fe(II) form,
40 which is readily available to phytoplankton. The photoreduction reaction
41 that produces Fe(II) in aerosols also produces the hydroxyl radical, which is
42 required for the oxidation of atmospheric DMS to MSA, and ultimately the
43 formation of CCN.

1 The fact that atmospheric Fe fluxes appear to play an important role in
ecosystem dynamics in many locations underscores the interwoven nature
3 of the links between climate change, the biogeochemical cycles of carbon,
nitrogen, and sulfur, and the potential for the oceans to sequester carbon.
5

7 **Changes in Oceanic Community Composition by Climate Changes**

9

11 Much of the literature on future climate change has ignored the possible
ecological shifts and changes in the pelagic food web that may occur as the
oceans warm. These changes may have a strong and significant impact on
13 the feedbacks between ocean biology and the physical climate system. Be-
cause much of the expected change is dictated by what community is present
and where in the world the community is located, there are few universal
15 answers (Kennedy et al., 2002; Poff et al., 2002). The most important climate
forcings are direct thermal effects (e.g., temperature-induced changes in
17 metabolism), and indirect, thermal effects (e.g., changes caused by ther-
mally induced changes to the environment such as decreased upwelling).
19

21 Paleoecological studies can give us a clue to the potential for future
climate-change-induced ecological shifts. For example, the glacial iron hy-
pothesis conjectures that an increase in aeolian Fe deposition to the South-
23 ern Ocean during the last glacial maximum (LGM) stimulated primary and
export production (Kumar et al., 1995) resulting in a decrease in atmos-
25 pheric $p\text{CO}_2$ (Martin, 1990). Evidence suggests that diatoms and co-
ccolithophorids did not play a prominent role in this increased production
27 (Howard and Prell, 1994; De La Rocha et al., 1998). One hypothesis that can
explain the increased production, as well as evidence for higher levels of
29 DMS-derived MSA, in ice cores during glacial times (Legrand et al., 1991), is
that the algal bloom-forming species responsible for the atmospheric $p\text{CO}_2$
31 drawdown was a high DMSP-producing organism that has left no sedimen-
tary record (DiTullio et al., 2000; Moore et al., 2000). Increased abundance of
33 the colonial haptophyte *P. antarctica* (a high DMSP producer) during the
LGM would be consistent with the accumulated evidence.

35 Expectations that current temperature levels will rise in the coming
decades can be detrimental or favorable to specific members of the ocean's
37 communities. Since there is a limit to the temperature rise in tropical waters
(based on limitations imposed by evaporative cooling), larger organismal
39 responses will occur in temperate and higher latitude regions. Temperature
changes will directly affect an organism's metabolism, growth, and fecun-
41 dity (among other things) and can be considered stressful as the temper-
ature moves outside the "normal" range for the organism. If the
43 temperature change is slow, it is possible that the organisms can shift their

1 distributional patterns to accommodate. In the case of dramatic changes in
2 temperature or where organisms cannot relocate to more acceptable climates,
3 the result can be reduced numbers or localized extinctions. Kennedy and
4 Mihursky (1971) demonstrated this in the laboratory using estuarine in-
5 vertebrates subjected to a short-term increase of 1 °C. In their experiments,
6 mortality increased from 0%, at the lower temperature, to 100% at the
7 higher.

8 Temperature increases can have positive effects related to increased fe-
9 cundity with temperature. In this case, some organisms will experience a
10 muting of seasonal declines in fecundity/abundance or year-round fecundity/
11 abundance increases. In both cases, the result will be a general annual mean
12 increase in numbers (assuming the temperature increase is sublethal). Of
13 course, changes in abundance of specific species, caused by migration or
14 increased fecundity/mortality, can fundamentally change the trophic inter-
15 dependencies in a particular region, leading to a change in the composition
16 of the underlying community. Terming the increase in abundance “positive”
17 may only make sense in a practical way for commercially important species.

18 The temperature tolerance range of many species may delimit their ge-
19 ographical distribution, with populations shifting latitudinally in response
20 to shifting climatic zones. For example, recent large-scale changes in the
21 biogeography of zooplankton (specifically, calanoid copepods) have been de-
22 tected in the northeastern North Atlantic Ocean (Beaugrand et al., 2002;
23 Beaugrand and Reid, 2003). Strong biogeographical shifts in all copepod
24 assemblages were found with a northward extension of more than 10°
25 latitude of warm-water species associated with a decrease in the number of
26 cold-water species. These changes were attributed to regional increase in
27 sea surface temperature.

28 Open ocean regions may be affected by changes in the thermohaline
29 pump, in which temperature-induced changes in density lead to mixing via
30 overturning of cold dense surface waters with less dense subsurface waters.
31 This mechanism is responsible for transport of oxygen and nutrients to the
32 deep ocean and would adversely impact organisms and communities in the
33 deep ocean. Carbon dioxide is also transported to the deep ocean, but a
34 slowdown would not affect deep-ocean organisms directly because there is
35 no light for CO₂ photosynthesis and the pressures at depth disallow forma-
36 tion of carbonate structures. However, some suggest that decreasing trans-
37 port to depth would increase the amount of CO₂ at the surface, decrease CO₂
38 uptake by the ocean from the atmosphere, and exacerbate CO₂ buildup in
39 the atmosphere. This is offset by those who believe that global warming will
40 lead to increased stratification of the surface waters, which would trap more
41 autotrophs at the surface causing an increase in photosynthesis and uptake
42 of CO₂, which in turn would lead to uptake by the ocean of atmospheric CO₂,
43 and a decrease in atmospheric CO₂.

1 Several lines of evidence from the Hawaii Ocean Time Series and the
2 limited historical data suggest that N₂ fixation is an important source of
3 new nitrogen for the open ocean ecosystems of the North Pacific Ocean (Karl
4 et al., 2001). These independent measurements and estimates during the
5 past couple of decades reveal increases in *Trichodesmium* (the dominate N₂
6 fixer) population abundances, increases in the molar N:P ratio above the
7 thermocline, and DIC drawdown in absence of nitrate and other forms of
8 fixed nitrogen. The nitrogen budget based on the data estimates and a one-
9 dimensional model calculation suggest that N₂ fixation presently supplies
10 up to half of the nitrogen required to sustain particulate material export
11 from the euphotic zone, but this source of new nitrogen from the N₂ fixation
12 process was much smaller before the 1980s. The increase in *Trichodesmium*
13 abundances and the high percentage of N₂-supported primary production
14 indicate that the ocean ecosystems are not in steady state. The changing
15 ocean ecosystems are influenced by either oceanic variability, or a new
16 quasi-steady state established in response to the Pacific decadal variability
17 (Karl, 1999).

18 Changes in ocean circulation affects many aspects of ecosystems, in-
19 cluding such things as nutrient distribution patterns (via upwelling and
20 horizontal transport) and the transport of larvae and juveniles by currents.
21 If the location of flow changes or the timing of that location changes then
22 species may not arrive at spawning or nurturing grounds at optimal times
23 and losses can occur to the species itself or to grazer/predator populations
24 relying on that species. Seasonal timings between predator and prey species
25 can also be affected. Many zooplanktons spend winter at depth and rise to
26 the surface waters in the spring. Depending on the community the phyto-
27 plankton may bloom at altered times, changing the conditions which lead to
28 zooplankton growth when they reach the surface waters.

29 Contemporary ecological data indicate that planktonic populations can
30 respond extremely sensitively and quickly to ocean variability. Long-term
31 climate–plankton connections have been detected in the Pacific in the Ca-
32 lCOFI program (e.g., Roemmich and McGowan, 1995) and in the North
33 Atlantic in the continuous plankton recorder (CPR) program (Colebrook,
34 1979). Phenological (seasonal) changes are also evident in the North Atlan-
35 tic CPR data, with some species reaching their seasonal peak up to 2 months
36 earlier in the 1990s compared to the long-term seasonal mean. The effects of
37 climate on plankton can take place on a worldwide scale and may be trans-
38 ferred from plankton to higher trophic levels; for example, by fish or bird
39 populations (Aebischer et al., 1990; Veit et al., 1996).

40 Analysis of long-term changes in phytoplankton, zooplankton, and
41 salmon in relation to hydrometeorological forcing in the northeast Atlantic
42 Ocean revealed significant relationships between (1) long-term changes in
43 all three trophic levels, (2) sea surface temperature in the northeastern

1 Atlantic, (3) NH temperature, and (4) the North Atlantic oscillation. The
2 similarities detected between plankton, salmon, temperature, and other
3 climatic parameters are also seen in their cyclical variability and in a step-
4 wise shift that started after a pronounced increase in NH temperature
5 anomalies at the end of the 1970s. Importantly, the changes flowed through
6 the entire food web, with all biological variables showing a pronounced
7 change over a relatively short time. These changes started after *circa* 1982
8 with a decline in euphausiids, followed by an increase in total abundance of
9 small copepods, an increase in phytoplankton biomass (1984), a decrease in
10 the large zooplankter *Calanus finmarchicus*, and by 1988, a decrease in
11 salmon. It is interesting to note that such a scenario of a decline in her-
12 bivores and top predators has also been found in laboratory microcosm ex-
13 periments which simulated the impact of global warming on an aquatic
14 ecosystem (Petchey et al., 1999.).

15 Coastal regions are affected by warming in a number of ways including
16 changes in precipitation/runoff, flooding, changes in salinity and oxygen
17 content, and changes in circulation patterns. Precipitation changes affect
18 runoff, which directly affects nutrient concentrations in the estuarine and
19 nearshore coastal areas. In addition, runoff is also a primary source of pol-
20 lutants in these regions. Increased runoff can inoculate coastal runoffs
21 leading to increased blooms of phytoplankton. These blooms, in turn can
22 affect the communities by removing needed nutrients, upsetting the oxygen
23 balance, introducing toxins, and affecting the depth that light penetrates to.
24 Not all species are capable of taking advantage of a sudden influx of nu-
25 trients and these events tend to favor the larger phytoplankton whose large
26 surface area can absorb proportionally more nutrients. Because of this, un-
27 expected inputs of nutrients can lead to localized changes in the community
28 composition in short order. Timing and seasonality of community cycles can
29 consequently be disrupted.

30 Coral reefs exist in a balance between water clarity and light penetra-
31 tion and occur at depths where the light penetration is optimal for growth of
32 the coral's zooxanthellae symbionts. Should any climate-related change
33 translate to a decrease in light penetration then the coral, and their reefs,
34 will be jeopardized. Increases in CO₂ can change the carbonate chemistry in
35 the ocean, which can lead to decreases in the amount of carbonate dissolved
36 in the water. Since carbonate is the principal component in "hard" corals, an
37 increase in CO₂ will, and already has, lead to destruction of the coral reefs.
38 The coral reef community then is destroyed not only by loss of the organisms
39 but by loss of habitat in the form of the reefs laid down by those organisms.

40 Changes in biodiversity, physiology, phenology, and geographic distri-
41 butions of plankton will likely alter competitive interactions between spe-
42 cies and trophic levels and may radically affect the marine food web, sea-to-
43 atmosphere carbon fluxes, and nutrient recycling processes. However, as

1 pointed out by Falkowski et al. (1998), our present lack of mechanistic un-
2 derstanding of the multiple feedbacks between marine ecosystems and cli-
3 mate limits our ability to make quantitative predictions.

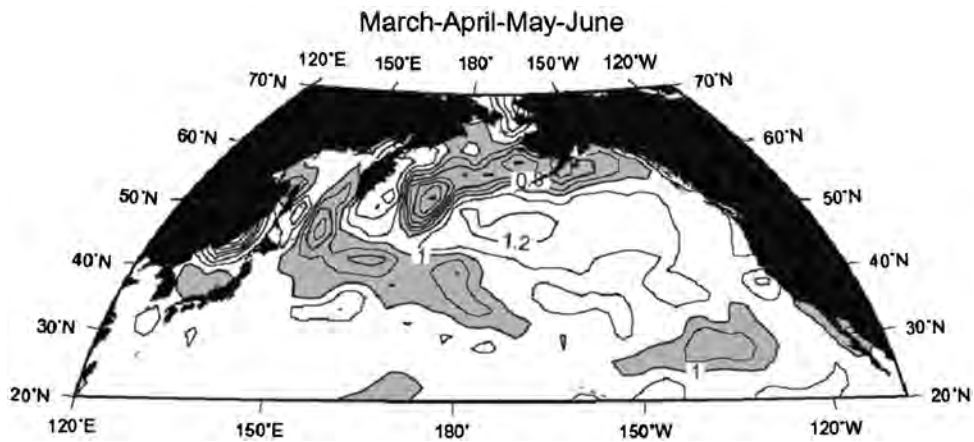
5 **8 Primary Productivity Response to Climate** 7 **Change in the North Pacific**

9 The North Pacific is of special interest because of its societal importance in
10 supporting fisheries and because of the consequent effect of the downstream
11 response over North America to possible feedbacks involving the oceanic
12 biology. Changes in the physical environment of the North Pacific driven by
13 global climate change can be expected to have an important effect on the
14 primary productivity and the entire local ecosystems.

15 North Pacific primary productivity changes in response to global warm-
16 ing have been examined by driving a diagnostic NPZ model by the envi-
17 ronmental changes predicted by a global ocean-atmosphere-coupled general
18 circulation model (O-A GCM), as described in detail by Pierce (2003). The
19 physical environment parameters that affect the biological model are MLD,
20 solar insolation, water temperature, and Ekman upwelling at the base of the
21 mixed layer. Although feedbacks to the physical climate are not included in
22 this run, it provides a remarkable view of how the ocean biology may change
23 in the globally warmed world.

24 The physical parameters that drive the ecosystem model were obtained
25 from a climate change projection run of the parallel climate model (PCM)
26 (Washington et al., 2000). This uses the CCM3 global atmospheric general
27 circulation model run at T42 resolution (approximately 2.8° in latitude and
28 longitude), coupled to the parallel ocean program (POP) ocean model run at
29 $2/3^\circ$ resolution. The coupled model includes land surface, runoff, and sea ice
30 components, and is forced by a so-called “business as usual” (IS92a) scenario
31 of future CO_2 and sulfur aerosol emissions. The PCM is relatively insen-
32 sitive to CO_2 forcing (e.g., Allen et al., 2001); most other major coupled
33 climate models show larger global temperature increases by the year 2100
34 than does PCM.

35 A result of this simulation is illustrated in Fig. 8, which shows the ratio
36 of primary productivity in the decade of the 2090s over the decade of the
37 2000s. Growing season values (March through June) are used here (but see
38 below). The main effect of the changes in physical environment is to de-
39 crease the primary productivity in the shaded regions of Fig. 8. Detailed
40 analysis (Pierce, 2003) shows that these changes are forced primarily by
41 increased stratification (a consequence of the warmer surface temperatures)
42 leading to a decline in MLDs during the winter. The shallower MLDs keep
43 the NPZ system closer to equilibrium, with a consequent reduction in the



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Figure 8: Ratio of primary productivity in the decade of the 2090s over the decade of the 2000s for the March through June growing season (Pierce, 2003).

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amplitude of the spring bloom. In other words, in the shaded regions of Fig. 8, the decade of the 2000s has deep wintertime-mixed layers that are associated with a sharp wintertime drop in phytoplankton (and consequently zooplankton) concentrations. This is partly due to mixing a given quantity of phytoplankton over a deeper layer and partly due to the lower average illumination levels in a deeper mixed layer. The rapid restratification of the water column in spring holds the phytoplankton near the surface in a well-illuminated region, and the depleted zooplankton cannot graze sufficiently quickly to prevent a spring phytoplankton bloom.

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In the 2090s, the shallower mixed layers keep the phytoplankton and zooplankton populations more evenly populated year round, with, as a result, less of a spring bloom. It follows that productivity during the rest of the year is somewhat higher in the shaded regions during the 2090s as compared to the 2000s, but this is generally not enough to overcome the loss of the spring bloom, resulting in a net yearly reduction of productivity over the majority of the shaded region in Fig. 8. In the nonshaded regions of Fig. 8, the general increase in temperature by the 2090s tends to increase primary productivity. The overall result, then, is a combination of a modest, near-uniform increase in productivity due to the warmer water combined with a sharp loss of springtime productivity in the regions where warmer surface waters cap the wintertime mixed layer in the future.

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It should be kept in mind that this line of analysis does not include other effects that will likely be important in the North Pacific, such as changes in the relative number of species, some of which might be better adapted than others to the changing environmental conditions. Also, future changes in the

1 biogeochemical environment (such as increased iron deposition from industrial activity in east Asia) could have a strong effect as well.

3 This type of coarse resolution climate model is unable to resolve the
4 dynamics of ocean boundary current systems. Yet primary production in
5 these boundary currents, e.g., in the eastern boundary upwelling system of
6 the North Pacific, is an important contributor to the earth's carbon budget.
7 Bakun (1990) suggested, based on observational evidence, that the warming
8 of ocean temperatures associated with greenhouse conditions would lead to
9 an inhibition of nighttime cooling and enhancement of daytime heating near
10 the coast. This leads to an intensification of the continental thermal lows
11 adjacent to the upwelling systems. The increase in onshore-offshore atmospheric
12 pressure gradient would then be translated into an intensification of
13 the coastal upwelling winds. More recently this idea has gained observational
14 and modeling support. Schwing and Mendelssohn (1997) report a
15 strengthening of upwelling favorable winds along the North Pacific eastern
16 boundary current. Snyder et al. (2003) find a significant increase in up-
17 welling favorable winds in a high-resolution regional climate model simulation
18 forced by greenhouse gases.

19 In contrast to these observations of increased upwelling favorable winds,
20 coastal observations over the last 50 years in the California Current System
21 reveal a transition towards conditions that are more typical of reduced up-
22 welling, such as a freshening of the surface waters in the coastal upwelling
23 boundary (Bograd and Lynn, 2003; Di Lorenzo et al., 2005) and a decline in
24 macrozooplankton abundance (Roemmich and McGowan, 1995). A possible
25 explanation for these seemingly contradictory lines of evidence involves the
26 observed increase in upper ocean stratification associated with warmer
27 temperatures. In this scenario, the stratification exerts a stronger control
28 than the winds on the ability of upwelling to supply subsurface nutrient-rich
29 water at the coast (McGowan et al., 2003). This hypothesis has been tested
30 with an eddy-resolving model of the coastal ocean driving a prognostic NPZ
31 ecosystem model (Fig. 9). Di Lorenzo et al. (2005) performed model exper-
32 iments that included as forcing conditions both the observed strengthening
33 of the upwelling winds and the warming trend over the last 50 years. The
34 effect of the increased stratification is strong enough in these experiments to
35 inhibit the otherwise upwelling favorable conditions. The model chlorophyll
36 response indicated a reduction in primary production in response to the
37 combined effects of upper-ocean heating and increased upwelling favorable
38 winds (Fig. 9).

39 Chai et al. (2003) used a coarse-resolution ocean model hindcast to de-
40 termine that decadal climate variability has the largest impact on oceanic
41 biological variability in the central North Pacific, a region bounded by two
42 oceanographic fronts at approximately 30–32° N (Subtropical Front) and
43 42–45° N (Subarctic Front) in the central Pacific. This was based upon their

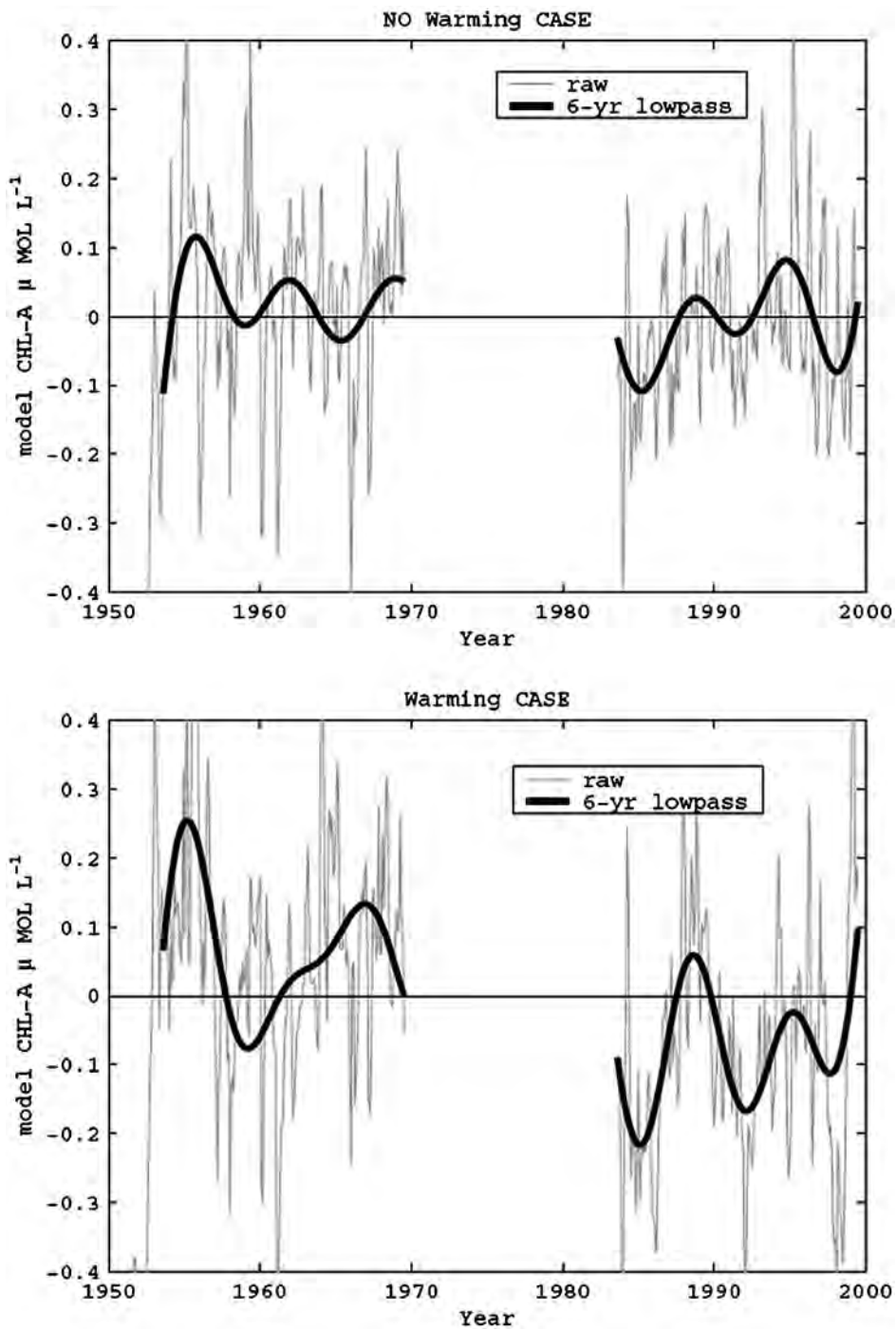


Figure 9: Time series of model surface Chl-a averaged over the eddy-resolving California Current model coastal boundary within 50 km from the coast. (a) Model experiment that include the strengthening of the upwelling winds but no warming conditions. (b) Same as (a) but the warming conditions are also included as forcing. See Di Lorenzo et al. (2005) for details of the physical model experiments.

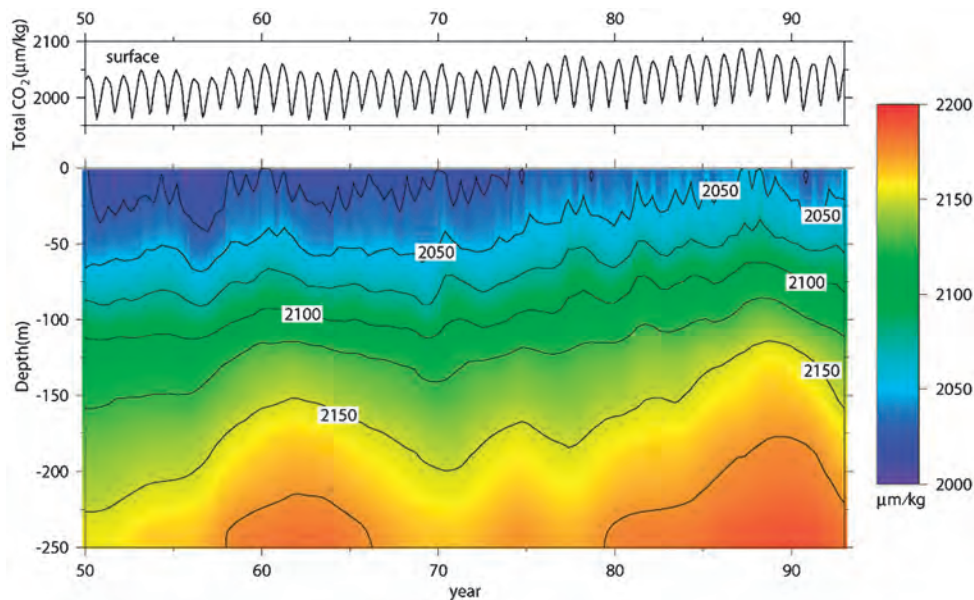


Figure 10: Time series of TCO_2 (unit: $\mu\text{mol kg}^{-1}$) in central north Pacific ($35^\circ\text{--}45^\circ\text{ N}$, $170^\circ\text{--}150^\circ\text{ W}$). Top panel shows the time series of surface TCO_2 concentration. The lower panel shows the vertical profile of modeled TCO_2 concentration from the surface to 250 m. The contour interval is $25 \mu\text{mol kg}^{-1}$.

analysis of the response of the MLD and the Ekman pumping to Pacific decadal variability. To extend this analysis and depict the possible changes in the carbon budget, Fig. 10 shows the spatially averaged modeled TCO_2 concentration for the central North Pacific (defined as follow: $35^\circ\text{ N--}45^\circ\text{ N}$, $170^\circ\text{ E--}150^\circ\text{ W}$).

The modeled monthly averaged TCO_2 concentration in the central North Pacific shows several scales of temporal variability between 1950 and 1993 (Fig. 10). First, the TCO_2 has a strong seasonal cycle in the central North Pacific, which is due to the seasonal cycle of upper-ocean physical conditions and biological uptake. The second most pronounced temporal variability between 1950 and 1993 is the influence of Pacific decadal variability. For example, the modeled TCO_2 concentration is higher during the 1980s and early 1960s, and the values are lower during 1970s and late 1960s. One of the reasons for the increase in modeled TCO_2 concentration during the 1980s is the change in ocean circulation and MLD in response to the wind pattern changes in the central North Pacific. Chai et al. (2003) found that the modeled winter MLD shows the largest increase between 30° N and 40° N in the central North Pacific (150° to 180° E), with a value 40–60% higher (deeper mixed layer) during 1979–1990 relative to 1964–1975 values. They

1 also found that the winter and annual mean Ekman pumping velocity
2 difference between 1979–1990 and 1964–1975 shows the largest increase
3 located between 30° N and 45° N in the central and eastern North Pacific
4 (180° to 150° W).

5 Beside the impact of decadal climate variability on the modeled TCO₂
6 concentration, there is another temporal trend (increasing from 1950 to
7 1993) in the modeled TCO₂ concentration in the central North Pacific. This
8 increase is due to anthropogenic effects because the model is forced with the
9 observed atmospheric *p*CO₂ from 1950 to 1993, which increased signifi-
10 cantly during this period. In order to separate the anthropogenic uptake and
11 storage of CO₂ from the natural cycle of climate impact, Chai et al. (in
12 preparation, 2005) conducted a second experiment in which the atmospheric
13 *p*CO₂ is fixed at 1950 values and other surface forcing are unchanged. By
14 comparing the results from this twin experiments, Chai et al. (2005) esti-
15 mated the anthropogenic uptake and storage of CO₂ in the Pacific Ocean
16 solely due to the changing of atmospheric *p*CO₂, eliminating the effects of
17 changing upper ocean circulation and mixing. The modeled anthropogenic
18 CO₂ has a linear trend since 1950 with a surface-increasing rate of
19 0.57 μmolkg⁻¹y⁻¹, which agrees with several independent estimates based on
20 the observations (Sabine et al., 2004).

QA :1

23 9 Conclusion

25 We have summarized many aspects of our current understanding of how
26 climate change due to increasing greenhouse gases will affect oceanic bi-
27 ology and how the physical–biological feedbacks may influence the evolving
28 physical climate system. The primary effects of ocean biology on physical
29 climate were its influence on the carbon cycle, the influence of oceanic
30 phytoplankton on upper-ocean absorption, and the influence of DMS pro-
31 duction by phytoplankton on atmospheric aerosols. The primary influences
32 of the physical climate on the ocean biology were the influence of aeolian
33 dust deposition and the multitude of ways that community structure can be
34 altered. Our focus was on the tropical and midlatitude Pacific Ocean, al-
35 though results from other ocean basins were also noted.

37 The greatest need for building on our current understanding of these
38 processes is long-term physical and biological observations in the ocean-
39 atmosphere system. Modeling efforts must be constrained and instigated by
40 these observational programs, and the observational strategies must be
41 motivated by the model results as well. Since we still lack an adequate
42 depiction of present-day oceanic community structures, it is very difficult to
43 determine how they will change over time and how the climate feedbacks
will be affected. Physical–biological modeling efforts should be directed to

1 those few oceanic regions where we currently have a fair understanding of
2 what species are present and how they vary naturally over time, such as the
3 southern California Current System (CalCOFI region). We can then assess
4 our skills at and prediction based on data gathered in the coming months or
5 years. This will provide a fair assessment of our skills of predicting the
6 responses of oceanic biology and the physical climate feedbacks on the cen-
7 tennial timescale of global warming.

9 **Uncited References**

11 Chang and Dickey, 2004; Coale et al., 1996; Leck et al., 1900.

13 **Acknowledgments**

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