

Effects of Ocean Biology on the Penetrative Radiation in a Coupled Climate Model

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1. Abstract

The influence of phytoplankton on the seasonal cycle and the mean global climate is investigated in a fully coupled climate model. The control experiment uses a fixed attenuation depth for shortwave radiation, while the attenuation depth in the experiment with biology is derived from phytoplankton concentrations simulated with a marine biogeochemical model coupled online to the ocean model. Some of the changes in the upper ocean are similar to the results from previous studies which did not use interactive atmospheres, for example amplification of the seasonal cycle, warming in upwelling regions such as the equatorial Pacific and the Arabian Sea and reduction in sea-ice cover in the high latitudes. In addition, positive feedbacks within the climate system cause a global shift of the seasonal cycle. The onset of spring is about 2 weeks earlier, which results in a more realistic representation of the seasons. Feedback mechanisms, such as increased windstress and changes in the shortwave radiation, lead to significant warming in the mid latitudes in summer and to seasonal modifications of the overall warming in the equatorial Pacific. Temperature changes also occur over land where they are sometimes even larger than over the ocean. The strength of interannual SST variability is reduced by about 10-15% and phase locking to the annual cycle is improved. The ENSO spectral peak is broader than in the experiment without biology and the dominant ENSO period is increased to around 5 years. Also the skewness of ENSO variabil-

ity is slightly improved. All of these changes lead to the conclusion that the influence of marine biology on the radiative budget of the upper ocean should be considered in detailed simulations of the Earth's climate.

2. Introduction

About half of the solar energy is within the spectral range of 350 to 700 nm, where pure sea water is highly transparent. Marine phytoplankton absorb light within this spectral range and thereby modulate the heat flux in the upper ocean (Morel and Maritorena 1988, 2001). The magnitude and variability of the solar radiation flux through the upper ocean layers has been investigated for various ocean regions (Ohlmann et al. 2000; Ohlmann and Siegel 2000). Light absorption by phytoplankton has potential implication for biological and physical processes as well as for ocean-atmosphere interactions. Ocean transports are influenced by the redistribution of heat, especially the partitioning of heat between the mixed layer and the deep ocean. Temperature changes, in turn, alter the dynamical turbulence in the ocean and the mixed layer depth. Sea-air heat and moisture fluxes are regulated by the surface ocean temperature and marine biology is itself affected by dynamical and temperature changes. Enhanced stratification can lead to better growth conditions for phytoplankton, while changes in mixing and upwelling may alter the amount of nutrients available. Limited growth in deeper layers due to self-shading is another process that can be addressed in a coupled ecosystem model.

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Previous studies focused on the tropical oceans. Observational evidence for a significant role of phytoplankton concentration for the upper-ocean heat budget in the equatorial Pacific is reported by Siegel et al. (1995), Chavez et al. (1998) and Strutton and Chavez (2003). Murtugudde et al. (2002) have tested possible effects on the upper tropical ocean circulation in an ocean general circulation model (GCM). Nakamoto et al. (2000) and Nakamoto et al. (2001) investigate the response of a mixed-layer isopycnal ocean GCM to chlorophyll pigments in the Arabian Sea and the equatorial Pacific. In their experiments the ocean GCM is driven with an atmospheric forcing and possible ocean-atmosphere feedbacks are not included. A global ocean GCM is applied by Manizza et al. (2004) to look at the bio-optical feedbacks between phytoplankton, ocean dynamics and sea-ice. Possible atmospheric responses are discussed in Shell et al. (2003). They first run an ocean GCM to obtain the biologically induced modifications of the sea surface temperature (SST) and then run a global atmospheric simulation using the modified SST. Again, possible ocean-atmosphere feedbacks are not included. A highly simplified coupled atmosphere-ocean model is applied by Marzeion et al. (2004) and Timmermann and Jin (2002), to study the influence of ocean biology on the tropical climate. Ocean turbidity through depth-dependent attenuation of solar radiation is also investigated in a bulk-type mixed layer model by Kara et al. (2004). Potential influences of phytoplankton in the Pacific on long term climate variations are discussed by Miller (2003).

In this paper a fully coupled, state-of-the-art climate model is, for the first time, utilized to investigate the influence of marine phytoplankton on global climate and its variability. The climate model includes an ocean/sea-ice GCM (MPI-OM) coupled on-line to an atmospheric GCM (ECHAM5) using the OASIS coupler. Both models are briefly described in section 2. With this system we are able to investigate the impacts on the SST, the resulting atmospheric changes and the feedback to the ocean dynamics. After discussing the global patterns in section 3 we focus on regions of special interest in section 4. These regions include the equatorial Pacific and the Arabian Sea, as well as the temperate oceans. We also address temperature and precipitation changes over land. In section 5, we summarize the results and draw

some conclusions as to the importance of the radiative effect of marine phytoplankton on the global climate.

3. Model Description

a. *Ocean general circulation model*

The Max-Planck-Institute ocean model (MPI-OM) is a z-coordinate global general circulation model based on primitive equations for a hydrostatic Boussinesq-fluid formulated with a free surface. Advection is computed with a second order total variation diminishing (TVD) scheme (Sweby 1984). It includes parameterizations of sub grid-scale mixing processes like isopycnal diffusion of the thermohaline fields and eddy-induced tracer transport following Gent et al. (1995), and a bottom boundary layer slope convection scheme. The dynamic and thermodynamic sea ice model with viscous-plastic rheology and snow follows Hibler (1979). The model uses an orthogonal curvilinear C-grid with a formal resolution of 1.5° . In this setup, one pole is located over Greenland and the other over Antarctica. The horizontal resolution gradually varies between 15 km in the Arctic and about 184 km in the Tropics. It has 40 vertical levels with level thickness increasing with depth. Eight layers are within the upper 90 m and 20 are within the upper 600m. The time step is 80 minutes. Bathymetry was created by interpolation of the ETOPO-5 dataset (Data Announcement 88-MGG-02, Digital relief of the Surface of the Earth. NOAA, National Geophysical Data Center, Boulder, Colorado, 1988) to the model grid. For details on MPI-OM and the grid versions, see Marsland et al. (2003). A parameterization that accounts for the effect of ocean currents on surface windstress (Jungclaus et al. 2004, this issue) is also included.

b. *Biogeochemical model*

The Hamburg Oceanic Carbon Cycle Model (HAMOCC5) simulates marine biology and biogeochemical tracers in the oceanic water column and the sediment. HAMOCC5 is coupled on-line to the circulation and diffusion of the MPI-OM ocean model, running with the same vertical and horizontal resolution and time step. Biogeochemical tracers are transported and mixed with the ocean advection and mixing schemes. The

carbon chemistry is identical to the HAMOCC3 version of Maier-Reimer (1993). The ecosystem model is based on nutrients, phytoplankton, zooplankton and detritus (NPZD-type) as described by Six and Maier-Reimer (1996). In addition new elements such as nitrogen, dissolved iron and dust are accounted for, and new processes like denitrification and N-fixation, formation of calcium carbonate and opaline shells, DMS production, dissolved iron uptake and release by biogenic particles and dust deposition and sinking are implemented. The dust fields for the experiments in this paper are taken from a model simulation by Timmreck and Schulz (2004). The model contains a sediment module following Heinze et al. (2003), including opal, silt, organic carbon and calcium carbonate.

c. Atmospheric general circulation model

The atmospheric general circulation model ECHAM5 (Roeckner et al. 2003) has evolved from the model of the European Center for Medium-Range Weather Forecasts (ECMWF). ECHAM5 solves prognostic equations for vorticity, divergence, surface pressure and temperature expressed in terms of spherical harmonics with a triangular truncation. Water vapor, cloud liquid water, cloud ice and trace components are transported with a flux form semi-Lagrangian transport scheme (Lin and Rood 1996) on a Gaussian grid. The solar radiation scheme (Fouquart and Bonnel 1980) has 4 spectral bands, 1 for the visible and ultra-violet, and 3 for the near-infrared.

d. Atmosphere ocean coupling

The atmosphere and ocean models run on different grids and with different timesteps. They are coupled with the OASIS coupler (Valcke et al. 2003) with a coupling timestep of one day. The ocean passes the SST, sea ice concentration and thickness, snow depth and the surface velocities to the atmosphere through OASIS. The atmosphere uses these boundary conditions for one coupling step and then transfers the surface forcing fields through OASIS back to the ocean model. Required surface forcing fields are heat, freshwater and momentum fluxes and the 10 m windspeed.

e. Model initialization and experiment setup

Two experiments are described in this study. A control run with constant shortwave attenuation in the upper ocean ("blue ocean") and an experiment with interactive marine biology ("green ocean"). The control run ("blue ocean") is identical to the experiment with windstress correction (WSC) as described by Jungclaus et al. (2004) in this issue. The "green ocean" is started from the same initial conditions as the "blue ocean" experiment and is integrated for 125 years.

In the "blue ocean" experiment, solar radiation is absorbed with a constant e-folding depth of 11 m (1/e of the light is left after 11 m depth). In the "green ocean" the incoming radiation is equally split into two wavelength bands. The first half consists of ultraviolet (UV) and infrared (IR) light and is fully absorbed within the first layer of the ocean model. The remaining half (I_0) is absorbed as a function of depth (z) according to the following equation, with a term for pure water and a term which is linear to the phytoplankton concentration:

$$I_z = I_0 * e^{k_w * z} * e^{k_{chl} * z} \quad (1)$$

Analogously to Moore et al. (2002), the attenuation coefficient for pure water is chosen to be $k_w = 0.04 m^{-1}$ and the coefficient linear to the phytoplankton concentration is computed as $k_{chl} = 0.03 * [Chl] m^{-1}$, with the chlorophyll concentration given in $mg Chl l^{-1}$. Chlorophyll is computed from phytoplankton with a fixed chlorophyll to carbon ratio of 1:80. There are more complex and more complete ways to calculate chlorophyll concentrations (Cloern et al. 1995) and to describe the radiation distribution (Morel and Maritorena 2001). However, a prerequisite for this experiment is that the two simulations of the climate system are comparable. This requires that the mean states of the "green" and the "blue" ocean setup are not too different. Therefore we intentionally keep the description as simple and easy to control as possible. The average penetration depth of the "green ocean" is about equal to the penetration depth of the "blue ocean" experiment.

4. Results

a. Global oceanic patterns

The penetration depth reflects the average distribution of chlorophyll. Figure 1 shows a yearly averaged chlorophyll distribution from the SeaWiFS Satellite¹ and the simulated surface chlorophyll. The chlorophyll concentration of the model is determined from the modeled phytoplankton biomass multiplied by a fixed chlorophyll-to-carbon ratio of 1:80. The high chlorophyll concentrations in coastal areas cannot be reproduced by the model because shelf processes and riverine input of nutrients are not captured.

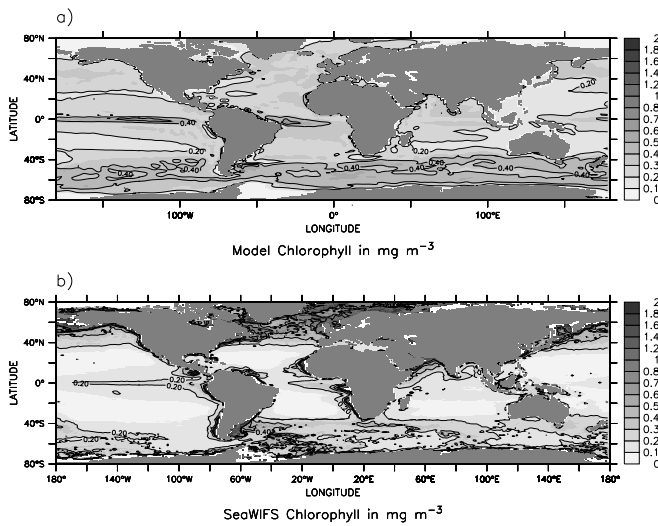


Figure 1: Annual averages of chlorophyll distribution from the SeaWiFS Satellite (above) and simulated by the model (below).

Averages of the resulting e-folding penetration depth ($k_w + k_{chl}$) are shown in figure 2. In the subtropical gyres the penetration depth is nearly identical to that of clear water all year round (25 m in our model), but in the

¹Data are provided by the SeaWiFS Project (Code 970.2) and the Distributed Active Archive Center (Code 902) at the Goddard Space Flight Center, Greenbelt, MD 20771. These activities are sponsored by NASA's Mission to Planet Earth Program.

upwelling regions at the equator it is always lower. In high latitude regions, penetration depth shows a strong seasonal cycle, with low penetration in their respective summers and high penetration in their respective winters. Also captured is the phytoplankton bloom due to upwelling during the summer monsoon in the Arabian Sea.

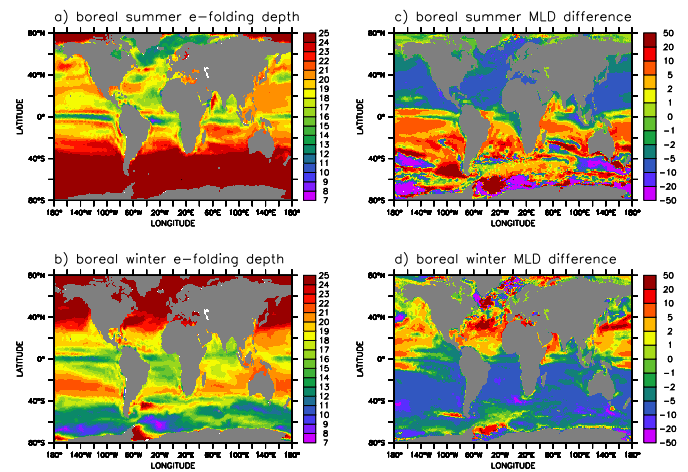


Figure 2: Left: 90 year average of the "green ocean" e-folding depth ($k_w + k_{chl}$) in meters. a) average e-folding depth during northern summer (June, July and August) and b) during northern winter (December, January and February). Right: 90 year average of the mixed layer depth (MLD) difference between the "green ocean" and the "blue ocean" experiment in meters. c) average MLD difference during northern summer (June, July and August) and d) during northern winter (December, January and February).

Heat flux in the upper ocean, modulated by phytoplankton, also affects the mixed layer depth (MLD). When most of the solar radiation is absorbed at the surface, surface waters warm up, the water column becomes more stable and the MLD shallows. Less cold, subsurface water is mixed to the surface and the surface water becomes even warmer. The converse is also true: When more solar radiation is able to penetrate into deeper layers, surface waters stay colder and the lower

layers heat up more. This leads to a destabilization of the water column and a deeper MLD. More cold, subsurface water is mixed to the surface and the surface water becomes even colder. Therefore the changes in the MLD act as an amplification of the effect of phytoplankton on SSTs that is larger than the direct warming effect. Differences in the temperate oceans are mostly around ± 5 m. The MLD of the "green ocean" is lower in summer and higher in winter compared to the "blue ocean". On annual average the MLDs are deeper in the subtropical gyres and shallower in the upwelling regions at the equator and along parts of the westerly coastline. Naturally, the differences are higher in winter at high latitudes where the MLDs are large. Much of these changes can be explained by changes in the sea-ice cover, although some are caused by changes in the wind fields. Shifts in the wind field patterns also cause the considerable differences in the "Roaring Forties" in the southern hemisphere during the austral winter (not shown).

Ocean-atmosphere interactions are determined by the sea surface temperature (SST). However, redistributing the heat in the surface ocean changes the SST, but also the temperature in the subsurface layers and the stability of the water column. A deepening of the ML also mixes warm surface waters into deeper layers. Temperature differences between the "green ocean" and the "blue ocean" experiment at the surface and down to 280 m are shown in figure 3. In contrast to the seasonal cycle at the surface, these large scale temperature change take decades to develop in the model. Therefore figure 3 shows an average of the last 40 years of the simulation. Temperature changes in the subsurface are larger than changes at the surface. The thermocline between 20°N and 20°S deepens and becomes steeper towards the subtropics (the deepening of the 20 degree isotherm depth is shown in contour lines in figure 3). Changes of up to 1°C reach as deep as 200 to 300 m between 40°N and 40°S , with the strongest warming in the Indian Ocean and the western Pacific. Temperature differences are at a maximum around 100 m between 20°N and 20°S . This is where the mixed layer has deepened and where warmer waters from the subtropics are transported to. Just underneath the surface at about 50 m, depth the warming concentrates at the upwelling regions underneath the equator and along the westerly coastline of Africa and South America. Through mixing and up-

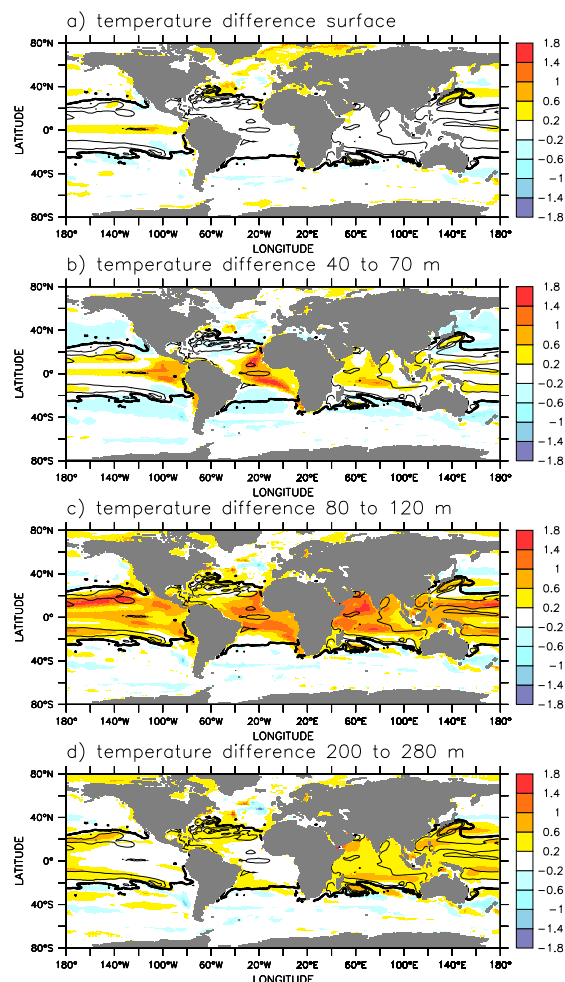


Figure 3: Annual average over the last 40 years of temperature differences between the "green ocean" and the "blue ocean" experiment in degrees celsius. a) is the SST, b) the average for the water column between 40 to 70 m, c) 80 to 120 m and d) 200 to 280m. Contour lines denote the deepening of the 20 degree isotherm depth between the two experiments in 10 m intervals.

welling the subsurface temperatures influence the temperatures at the surface. In this model, the warming of

the "cold tongue" in the equatorial Pacific is to a large part due to the deepening of the thermocline.

b. Global atmospheric patterns

The zonal mean temperature differences ("green" minus "blue" ocean) in figure 4 show that the influence of the SST changes extend through the troposphere. The dominant pattern is the amplification of the seasonal cycle. During the northern summer we find a warming north of 30°N and a cooling in the southern hemisphere, strongest around 60°S . During the northern winter this pattern partly reverses. Due to a retreat in sea ice the high latitudes are warmer in their respective winters. Air temperatures over the equatorial regions are warmer all year round. SSTs across the tropical Pacific contribute significantly to the observed patterns of tropical rainfall. In the "green ocean" experiment, rainfall is higher over the western tropical Pacific and slightly lower in the eastern equatorial Pacific. North and south of the equator there is a shift in the rainfall patterns. The wind systems across the tropical Pacific are associated with the equatorial patterns of SST and rainfall. In the "green ocean" experiment the Hadley cells become stronger and, in contrast to Shell et al. (2003), we also find a minor increase of the Walker circulation.

c. Seasonal changes

From the subtropics to the high latitudes we generally see an amplification of the seasonal cycle and the warming of the sea surface in spring starts about two weeks earlier. Temperature changes also occur over land and are often larger than over water. Figure 5 shows the temperature differences between the "green ocean" and the "blue ocean" for 4 seasons (northern spring to winter) and lines marking the 99% significance level. During northern spring and summer most of the warming occurs north of the subtropical gyres and extends well over land. Between 20°N and 60°N there is a cooling in winter that is even stronger over land. Along with the amplification we find a shift of the seasonal cycle by about 2 weeks. To illustrate the shift, figure 6 shows two timeseries of SSTs from 40°N in the Pacific and the Atlantic and 40°S in the Pacific. In the southern hemisphere the shift of the seasonal cycle is also of the order of two weeks, but the SST differences are smaller. The earlier start

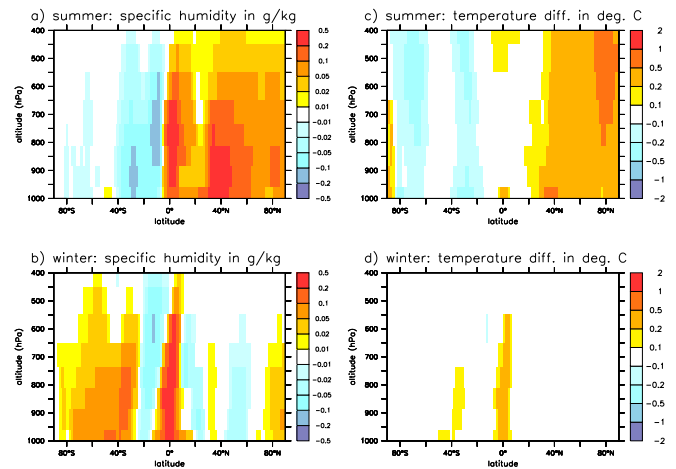


Figure 4: Zonal average of temperature and specific humidity over the Pacific. Shown are differences between the "green ocean" and the "blue ocean" experiment from the ground up to 400 hPa. A) and C) are averaged from June - August and B) and D) from December to February.

of the spring warming is directly caused by the influence of the phytoplankton spring bloom on the upper surface heating and the MLD. Comparison with SST data from the NCEP/NCAR reanalysis (Kalnay et al. 1996) shows that the shift makes the simulation of the seasonal cycles more realistic. This is an indication that phytoplankton has an influence of similar magnitude on the seasonal cycle in the real world.

Sea-ice in the Arctic and Antarctic is generally lower in the "green ocean" experiment, with stronger retreat in the north. This is associated with a warming of the SST in the high latitudes of about 0.4°C north of 60°N and 0.1°C south of 60°S , on the annual average. While summer SSTs are warmer in the Labrador Sea, along the coast of Greenland and in the Barents Sea, temperatures over the Arctic are higher in autumn and winter. This influences the land temperatures over Siberia and Alaska in the autumn.

The atmosphere is directly influenced by the ocean through the latent and sensible heat fluxes (which are

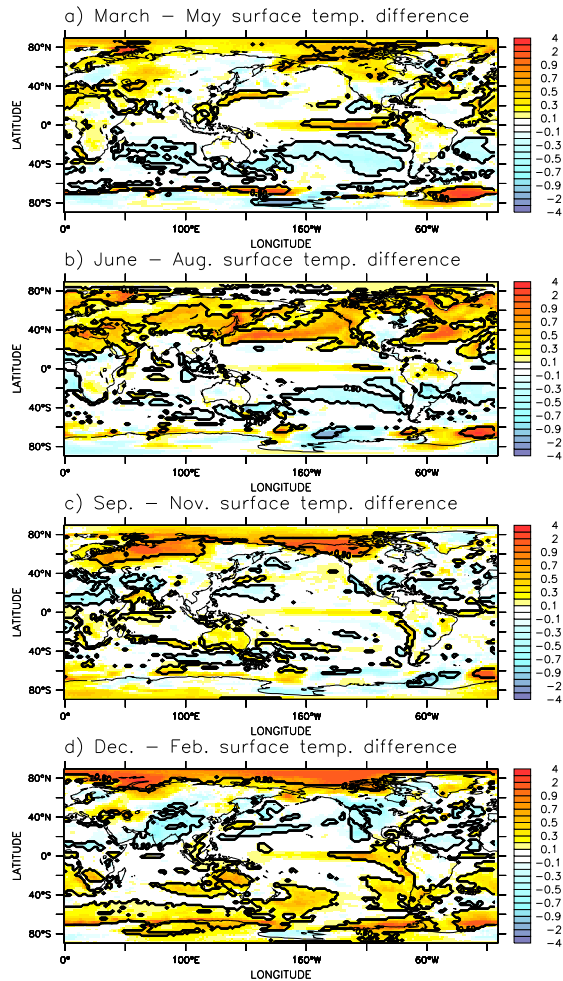


Figure 5: Seasonal averages of the last 60 years of 2 m surface temperature differences between the "green ocean" and the "blue ocean" experiment in degrees celsius. The contour lines mark the 99% significance level. a) is the average for March - May, b) is the average for June - August, c) is the average for September - August and d) is the average for November - February.

related to the SST). Most of the feedback is provided in the form of changes in windstress, shortwave radiation

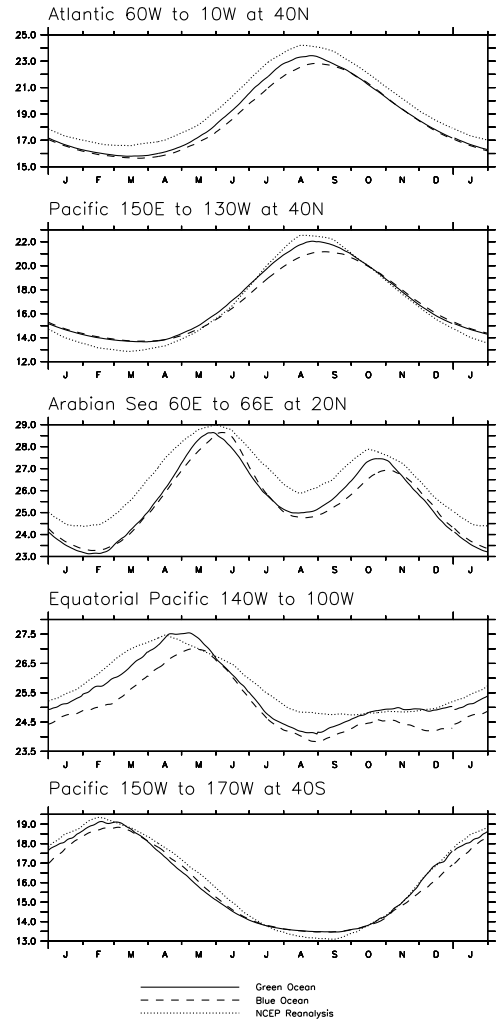


Figure 6: Seasonal SST cycle of the "green ocean" and the "blue ocean" experiment in the North Pacific, North Atlantic, Arabian Sea, Equatorial Pacific and South Pacific. Shown are 50 year averages of daily values. Also shown for comparison are climatological averages from the NCEP/NCAR reanalysis.

and precipitation. Heat flux and atmospheric feedbacks for the regions used in figure 6 can be seen in figure 7. Precipitation (results not shown) generally increases be-

tween 10°N to 10°S over the tropics and decreases between 30° to 10° over the subtropics. Precipitation also increases by 3 mm d⁻¹ along the coast of India when the winds blow over the Arabian Sea during the summer monsoon from July to September. Elsewhere the changes in precipitation are small. Another feedback is due to the shortwave radiation incident on the ocean. On average, shortwave radiation is up to 4% higher over the eastern equatorial Pacific and about 4% lower elsewhere in the tropics.

The strong warming north of 30°N during the northern summer is not replicated in the southern hemisphere during the southern summer (figure 5). The reason for this asymmetry is that the summer warming in the north is amplified by a reduction in cloud cover, resulting in an increase of shortwave radiation by about 10% (figure 7). The reduction in cloud cover can not be explained with local feedbacks, but with atmospheric teleconnections from the SST anomaly in the equatorial Pacific. Figure 5 shows that the SST difference there is high in spring (March, April and May) and low during the rest of the year. The teleconnection is based on a strengthening of the Hadley circulation with higher precipitation over the equator and anomalous drying and warming in the subtropical subsidence regions. This "atmospheric bridge" connects the equator and the extra-tropics with a time lag of about 2-3 month (Alexander et al. 2002), and explains the amplification of the marine biology induced warming in June, July and August. The anomalous warming north of 30°N in the Pacific and the Atlantic during the northern summer abruptly ends at the beginning of autumn (figure 7) when the windstress increases. Although the cloud cover north of 30°S over the southern hemisphere is reduced at the same time in the "green ocean", this does not result in warmer surface temperatures for two reasons. Firstly, there is less solar radiation during the southern winter and secondly, due to the enhanced seasonal cycle the surface temperatures are already lower in the southern hemisphere and the increase in shortwave radiation only moderates the cooling.

d. Regions of special Interest

i. Equatorial Pacific The region of upwelling along the equatorial Pacific is called the "cold tongue". In most state-of-the-art ocean GCMs the simulated cold tongue

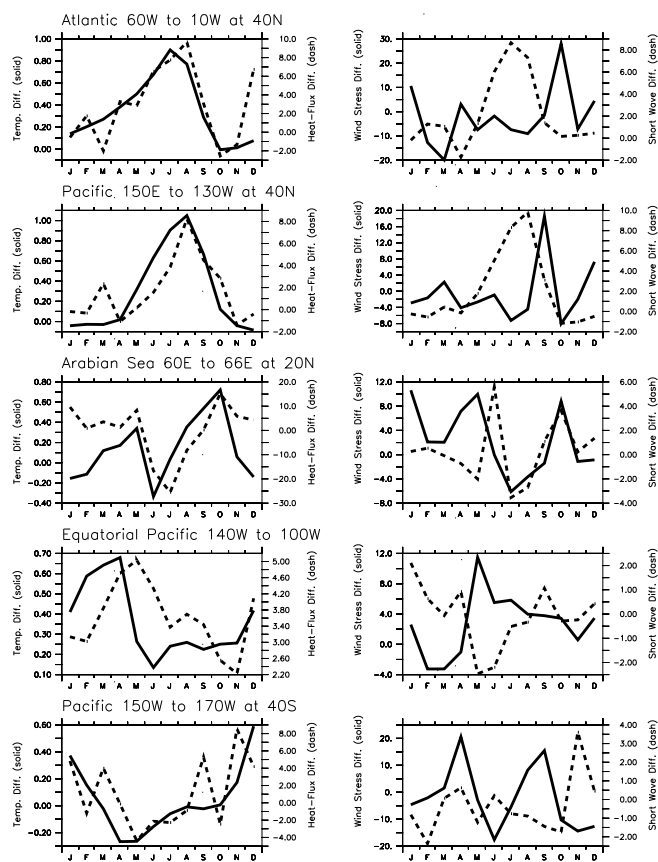


Figure 7: Left: solid line are are surface temperature differences in degree celsius. Dotted lines are the changes in the sensible and latent heat flux (ocean to atmosphere) in Wm⁻². Right: feedbacks from the atmosphere to the ocean. Solid lines are percentage windstress differences, dotted ones are shortwave radiation in Wm⁻². Shown are 90 year averages of the seasonal cycle from the "green ocean" and the "blue ocean" experiment in the North Pacific, North Atlantic, Arabian Sea, Equatorial Pacific and South Pacific.

is colder than observed; this phenomenon is called the "cold bias". In coupled ocean-atmosphere models the offset is usually larger than in forced simulations because

of coupled feedbacks. In addition, a cold anomaly in the eastern equatorial Pacific leads to an increase of the trade winds and amplifies the upwelling. This is known as the Bjerkness feedback. Taking into account the effect that surface currents have on wind shear has improved the model significantly (Jungclaus et al. 2004, this issue). However, the model still shows a "cold bias" of more than 1°C , mostly during the upwelling season from July to September. In the "green ocean" this bias improves by over 0.5°C on average (figure 3a). Figure 6 shows the seasonal cycle of SST differences in the equatorial Pacific. Most importantly, the spring warming has shifted by 2-3 weeks from May to April and the SST in April is up to 1.2°C warmer. The direct comparison with SSTs from the NCEP/NCAR reanalysis (Kalnay et al. 1996) shows that the shift of the spring warming and the increase in temperature is an improvement towards a more realistic representation of the seasonal cycle in the equatorial Pacific. During the upwelling season the "green ocean" is only slightly warmer than the "blue ocean" and both experiments are up to 1°C colder than the NCEP/NCAR reanalysis SST.

In contrast to our results, Murtugudde et al. (2002) find that effects of penetrative radiation are largest during the upwelling season. Their finding is in agreement with a sensitivity study (results not shown) conducted using the MPI-OM ocean GCM forced with surface fluxes from the NCEP/NCAR reanalysis (Kalnay et al. 1996). This is surprising, as the mechanisms that cause the warming in the equatorial Pacific are the same in both the stand alone ocean and the coupled ocean-atmosphere model. Most of the warming is caused by the deepening of the thermocline (figure 3). In addition, nutrients upwelled from below during the cold phases cause an increase in the phytoplankton biomass, which again traps the heat of the solar radiation at the surface. Atmospheric coupling reverses the seasonal pattern. In spring, SSTs are warmer at the equator and colder in the subtropics. This causes a strengthening of the trade-winds (figure 7) which leads to stronger upwelling and almost completely counteracts the warming mechanisms in the following month.

ii. Arabian Sea Accurate simulation of SSTs in the Arabian Sea is important because it influences the monsoon system over the region. The seasonal cycle of

the SST in the Arabian Sea has two maxima during the inter-monsoon periods and minima during the southwest summer monsoon (June - September) and during the northeast winter monsoon (December - March). Satellite observations from SeaWiFS show a chlorophyll peak during the onset of spring (February and March) and in the newly upwelled waters from July to September. The model captures the low chlorophyll concentrations observed by SeaWiFS from November to January and the rise at 20°N in spring, but it underestimates the boreal summer bloom (not shown). However, we find the largest differences between the "blue" and the "green" ocean experiments in autumn. Surface temperatures in the west are over 1°C warmer and the warming stretches to the east at 20°N . In addition, the warming after the summer monsoon starts almost one month earlier and the peak also shifts by 2-3 weeks. During the summer monsoon from July to September, when the winds blow over the Arabian Sea, precipitation along the coast of India increases by about 3 mm d^{-1} .

One would expect a large influence from the phytoplankton during the spring-bloom in February and March, but we find only a shift of the peak by 1-2 weeks and no additional warming. However, figure 6 indicates that both the warming and the shift lead to a representation of the seasonal cycle that is more similar to the SST from the NCEP/NCAR reanalysis (Kalnay et al. 1996). Figure 7 shows that the windstress during both monsoon seasons has also increased in the "green ocean" experiment. In this model, SST anomalies in autumn mostly result from the deeper thermocline in the "green ocean" experiment. The shifts in the seasonal cycle in spring and autumn can not be explained by local feedbacks, but only by the large scale shift in the seasonal cycle over the northern and southern hemispheres.

Our results are again quite different to the studies without atmospheric feedbacks by Murtugudde et al. (2002) and Nakamoto et al. (2000). They both find the maximum warming in March and April, a slight cooling from July to August and a minor warming from October through January.

e. Tropical Pacific interannual variability

In this section the impact of biology on the interannual variability of the tropical Pacific is described. For this

purpose only years 26 to 115 of both simulations are considered. The period is limited by the length of the "green ocean" simulation, and the first 25 years of the simulations are considered as part of the model spinup.

One impact of biology in the model is a reduction in variability. In the central eastern Pacific the standard deviation of the SST drops by around 15%, from 1.75° C to 1.5° C (figure 8). The observed value is somewhere between 1 and 1.5° C, depending on both the period considered and the dataset used. The structure of the variability is also reminiscent of the observations, except that it extends too far into the warm pool. Variability in the extra-tropics, under the storm track zones, is overestimated; the causes for this are being investigated. Compared to the control simulation, the global structure of the interannual variability remains quite similar.

It has been suggested that biological processes may contribute to the skewness of ENSO variability. In our simulation there are some hints that negative ENSO events are slightly more damped than the positive ones: in the biologically active simulation Niño3 (150°W -90°W, 5°S-5°N) averaged SST anomalies (SSTa) vary between -3° C and 3.5° C (figure 9b), whereas in the biologically inactive simulation Niño3 SST anomalies vary from -3.5° C to 3.5° C (figure 9c). Observed variability over the last 100 years (Rayner et al. 2003, HadISST) varies from around -2° C to almost 3.5° C (figure 9a). The frequency of strong positive events (greater than, say, 3° C) is also slightly reduced in the biologically active simulation, and is more consistent with the observations. The histogram of Niño3 SSTs from the "green ocean" simulation compares more favorably to the observed one than that of the simulation without biology (figure 10). Thus, biological process may positively affect the skewness of ENSO in our simulation, however these changes may well not be statistically significant given the shortness of our simulation.

The inclusion of biological processes also brings improvements to the seasonal variations in ENSO variability. The observed variations show the strongest variability in December and the weakest variability in April (figure 11). Both the biologically active and inactive simulations have significantly stronger variability than the observations in all calendar months (figure 11), although the biology run is always weaker than the control run. The differences are most pronounced in April, with the

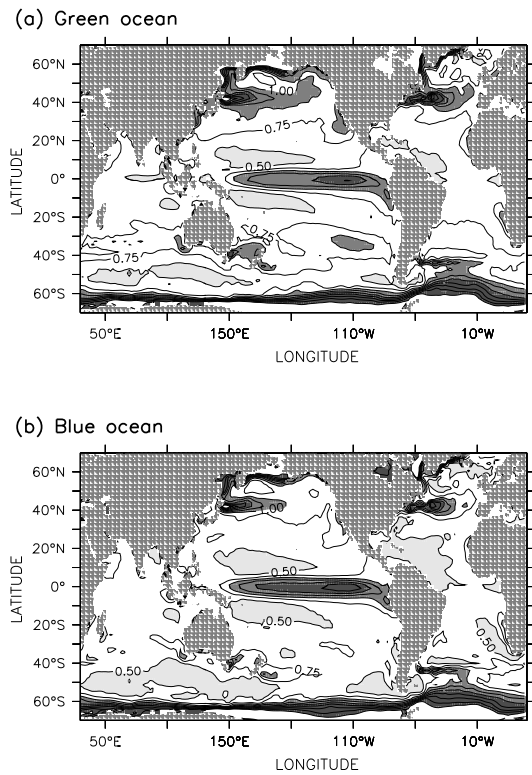


Figure 8: Standard deviation of SST anomalies from years 25 to 115 of (a) the biologically active (green ocean) and (b) biologically inactive (blue ocean) simulations. The contour interval is 0.5° C, with values less (larger) than 0.5° C (1.0° C) light (dark) shaded.

observed minimum in variability better represented in the green ocean. This is an indication that the termination of ENSO events is better represented in the biological simulation. The unrealistic second peak in variability that occurs in late boreal summer in both simulations is also less pronounced when biology is included. This second peak is mainly the result of the poor simulation of negative anomalies.

The inclusion of biological processes also improves the simulation in terms of the periodicity of interannual variability. The observed spectrum of Niño3 SSTa calculated for the time-span 1910 till 1999 shows a broad

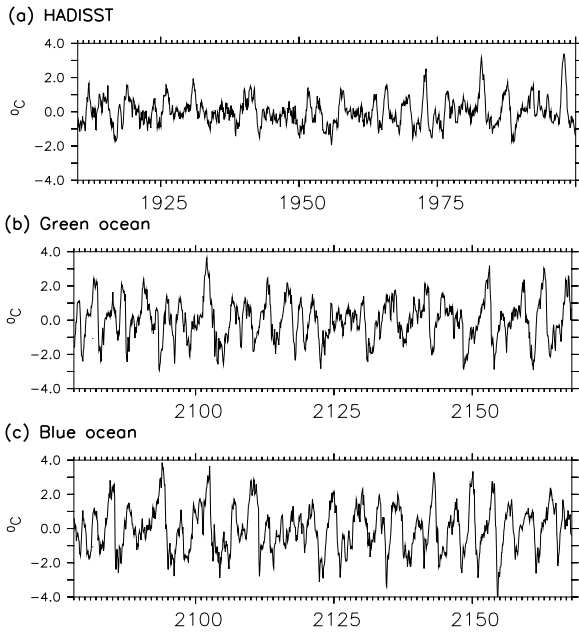


Figure 9: Niño3 averaged SST anomalies from (a) observations (HADISST) between 1910 and 1999, and years 25 to 125 of (b) the biologically active and (c) biologically inactive simulations

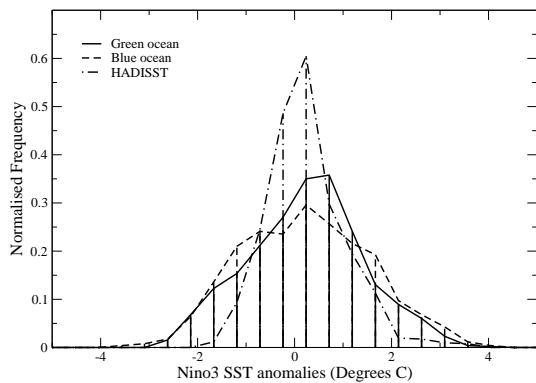


Figure 10: Histogram of the Niño3 averaged SST anomalies shown in figure 9.

peak between 3 and 5 years, and a second weaker bump around 2.5 years (figure 12). Only 90 years of observations are considered here to be consistent with the

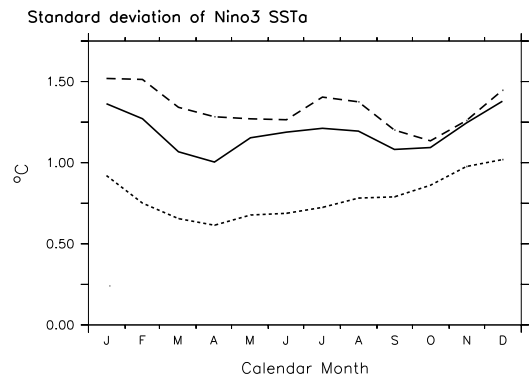


Figure 11: Standard deviation of Niño3 averaged SST anomalies as a function of calendar month for HADISST observations (1910-1999, dotted), and years 25 to 125 of the "green ocean" (solid) and "blue ocean" (dashed) simulations.

length of the simulations. The biological inactive simulation shows a sharp spectral peak at around 3.5 years, and a second weaker peak at 8 years (figure 12). When a longer period of the simulation is considered this second peak weakens, and the main peak shifts to around 4 year. The inclusion of biological processes causes the spectral peak to broaden, and the dominant period to shift to 5 years, in better agreement with the observations (figure 12). A weak bump at around 2.5 years is also simulated, but given the shortness of the timeseries it may not be a significant feature of the spectrum.

The propagation characteristics, and the spatial structure of SST anomalies simulated in the "green ocean" run are quite similar to that of the biologically inactive "blue ocean" simulation, and are only briefly described here (see article by Jungclaus et al. this issue). Simulated variability is weak west of 140°W, and strongest in the central and eastern Pacific, where anomalies range between ±4° C. These aspects compare favorably with observed SST variability, which is weak west of the date-line; in the central Pacific anomalies range between ±1-2° C, and in the east they range between between -2° C and 4° C. The westward extension of the anomalies and their intensity in central Pacific are overestimated (not shown). However, similar to the observations, SST anomalies tend to develop at almost the same time in the

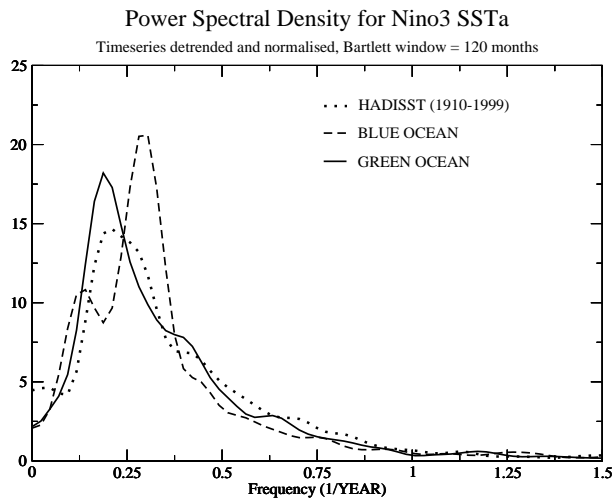


Figure 12: Power spectra of Niño3 averaged SST anomalies from HADISST observations (1910-1999), and years 25 to 125 of the "green ocean" and "blue ocean" simulations.

east and central Pacific. Also reminiscent of the observations, individual events appear to propagate sometimes weakly eastward or westward. The meridional extent of the ENSO-related SST anomalies is very similar in both simulations, and also similar to the observations (Jungclaus et al. 2004, this issue).

In summary, the inclusion of biological processes reduces the strength of interannual SST variability by about 10-15%, improves the skewness and phase locking to the annual cycle, increases the dominant ENSO period to around 5 years, and broadens the ENSO spectral peak. Changes in ENSO properties are caused by changes in the ENSO feedbacks, which in turn may result directly from changes in model physics or indirectly from changes in the mean state.

As discussed above, the inclusion of biological processes results in a warming of 0.5°C in the eastern Pacific. The warming is, however, not primarily due to a shallowing of the mixed layer, since close to the equator the mixed layer depth shallows by no more than 3 m. The warming appears primarily due to a deepening of the thermocline from 20°S - 20°N (about 10 m on average), which at the equator results in warmer water being upwelled. The warming effect is, however, somewhat

compensated for by a strengthening of the overturning of the subtropical cells, including equatorial upwelling, of about 10%. The changes in the subtropical cells and the deepening of the thermocline are driven by changes in the tropical windstress curl, and not by changes directly at the equator where the mean zonal winds are hardly affected. The Bjerkness feedback is apparently not excited in our model, since the induced SST anomalies are located too far east of the convection.

Despite the deepening of the thermocline, the thermocline SST feedback in the eastern Pacific (150°E - 90°W) was insignificantly changed. The damping effect of the atmosphere there is also hardly changed. In contrast, the sensitivity of the atmosphere to SST anomalies is increased back to levels close to those seen before the introduction of the wind shear correction (Jungclaus et al. 2004, this issue). This change would, however, argue for an increase in ENSO variability, in contrast to the decrease simulated.

The interaction between biological processes and ENSO are investigated to see if they could explain the simulated decrease in ENSO variability. In our simulation, the relationship between biological activity and upwelling is highly non-linear: the two are only related below a certain upwelling threshold. Below this threshold increases (decreases) in upwelling result in increases (decreases) in phytoplankton concentration and hence to a reduction (increase) in the optical attenuation depth. Above this threshold biological activity is controlled by other factors. Furthermore, upwelling and ENSO activity are only strongly related in the central Pacific in our model. Thus only in the central Pacific, and only during warm events, is there any coherent variability in phytoplankton concentration: a decrease in phytoplankton concentration and an increase in the optical penetration depth is seen. This may support a reduction in the strength of warm events, due to the direct effect of deeper penetration of solar radiation, but does not argue for any change in cold events. Our model simulates no coherent changes in mixed layer (also known as the indirect effect).

Analysis of the relevant ENSO feedbacks would thus suggest that the changes in the simulated ENSO properties are due to changes in the mean state. Indeed the theoretical work of Federov and Philander (2001) predicts that the deepening of the thermocline by about

10 m would result in a lengthening of the ENSO period of the order of the simulated change. In their work, a deepening of the thermocline results in a shift towards a more "thermocline mode". From our analysis it is, however, hard to confirm that this is indeed the case. A shift towards such a regime should also result in changes in ENSO amplitude, but such a relationship seems not to have been discussed in the literature.

Our results are in contrast to those of Marzeion et al. (2004). In their hybrid coupled modeling study, the inclusion of biological processes results in a decrease in the strength of the annual cycle, an increase in ENSO activity, and an increase in the dominant ENSO period (from 2 to 3 years). In their case, the strengthening of ENSO variability is due to the weakening of the annual cycle. Our results are consistent with theirs in so far as we get a lengthening of the ENSO period, and that they also show the importance of coupled interactions. However, our results differ from theirs in that we get very little change in the strength of the annual cycle, and our ENSO variability decreases in strength. Unlike them, in our model the Bjerkness feedback is not strongly excited, and the changes in ENSO properties are largely driven by changes in the off-equatorial winds. In our model the indirect effect on mixed layer depth is also not active.

5. Conclusions

We investigate the influence of phytoplankton on the seasonal cycle and the mean global climate in a fully coupled climate model. We primarily do this by analyzing the differences between a "green ocean" with phytoplankton and a "blue ocean" control experiment without biology. One has therefore always to keep in mind that the way radiation is treated in the control experiment is as important for the difference between the two experiments, as the effect of phytoplankton on the radiation. In this study, we use a fixed attenuation depth of 11 meters in the "blue ocean" control run and a "green ocean" setup where half of the radiation is absorbed in the first layer and the other half is attenuated with an attenuation depth that varies between 0 and 25 meters, dependent on biology. With these settings, the average attenuation depth in the "green ocean" is about equal to the fixed attenuation depth in the "blue ocean", and the mean states of

both systems are similar. This is a necessary prerequisite, as a fully coupled climate system has nonlinear feedbacks that would make two experiments with a different mean state very difficult to compare to each other.

Previous studies are all based on forced simulations, except those which use a simplified coupled system (Timmermann and Jin 2002; Marzeion et al. 2004). They therefore do not have such constraints and mostly use a "clear water" assumption (23 m attenuation depth) in their control experiments. This is an important factor to keep in mind when we compare our results with earlier studies. Murtugudde et al. (2002) refer to a control run that use an attenuation depth of 17 meters. They report that the warming of the water column below the mixed layer leads to a warming of SSTs in the equatorial Pacific cold tongue region of up to 1° C. This is in agreement with our findings. We find a even larger warm anomaly in the equatorial Pacific in the same model when we force it with surface fluxes from the NCEP/NCAR reanalysis (Kalnay et al. 1996) (data not shown). Similar to Murtugudde et al. (2002) we also see a warming in the upwelling regions of the Arabian Sea.

In contrast, Nakamoto et al. (2001) and Manizza et al. (2004) who compare the impact of phytoplankton to a control run with an attenuation depth of 23 meters and both find a cooling in the tropics. What appears to be in a contradiction is actually consistent. Most of the warming in the tropics is caused by a deepening of the thermocline, due to a higher attenuation depth in the subtropics. With a "clear water" assumption in the control experiment, biology reduces the attenuation depth almost everywhere and the thermocline becomes more shallow, which results in a cooling along the equator. In addition, the impact of phytoplankton on SST and MLD is substantially different in different types of ocean GCMs. Kara et al. (2004) use a control run attenuation depth of 17 meters in a bulk-type mixed layer model and only find minor impact of phytoplankton on the SST in the equatorial Pacific. Despite such differences, the direct influence of marine biology in our ocean model is similar to the results from previous studies. Like Manizza et al. (2004), we find that the seasonal cycle in the northern and southern hemisphere is amplified and there is less sea-ice in the high latitudes.

In addition to the amplification, we find a shift of the seasonal cycle by about 2 weeks. The earlier start of the

spring warming is directly caused by the influence of the phytoplankton spring bloom on the upper surface heating and the MLD. The comparison with SST data from the NCEP/NCAR reanalysis shows that the shift makes the simulation of the seasonal cycles more realistic. This is an indication that the phytoplankton in the real world could have an influence on the seasonal cycle of a similar magnitude. The shift and the warming are more pronounced in the northern hemisphere. In the southern hemisphere the seasonal cycle is also shifted by about two weeks, but the SST differences are smaller. We attribute some of the additional warming north of 30°N to an atmospheric teleconnection from the tropics. SST differences between the "blue" and the "green" ocean in the equatorial Pacific are not uniform over the year. The warm anomaly is high in spring and lowered by higher windstress curl during the rest of the year. The "atmospheric bridge" that connects the equator and the extratropics has a time lag of about a season (Alexander et al. 2002), which explains the amplification of the warming during the northern summer.

The increase in the windstress curl along the equator is due to the cooling in the subtropics and the warming in the tropics. It leads to a average strengthening of the equatorial upwelling by about 10% that is strongest during the northern summer.

A La Niña state is amplified by the Bjerkness feedback; a cold anomaly in the eastern equatorial Pacific leads to an increase in the strength of the trade winds. This leads to more upwelling and even colder temperatures. As discussed by Timmermann and Jin (2002), biologically induced surface heating can amplify the warming by an inverse Bjerkness feedback. In our model the Bjerkness feedback is not strongly excited, since the induced SST anomalies are too far east of the convection. Compared to a "clear water" (23 m attenuation depth) experiment, Timmermann and Jin (2002) find an additional heating of 0.55° C/month from heat absorption by phytoplankton and the indirect effect of the mixed layer depth. In our experiment the MLD shallows by no more than 3 m and the direct heat absorption is also of minor importance.

In our experiment, we find a reduction in the strength of interannual SST variability by about 10-15%. The skewness and the phase locking to the annual cycle are also improved. The ENSO spectral peak is broader than in the "blue ocean" and the dominant ENSO period is in-

creased to around 5 years.

Like Shell et al. (2003), we find that temperature changes also occur over land, where they are often larger than over the ocean. The warming in summer and the change in the seasonal cycle influence large parts of the northern hemisphere. There is also a significant cooling over parts of Asia and North America in winter. Changes in sea-ice influence large areas in Siberia and Alaska in autumn.

All of these changes lead to the conclusion that the influence of marine biology on the radiative budget of the upper ocean should be considered in detailed simulations of the earth's climate. It is a part of the climate system that is important for the seasonal cycle. There are also indications in this experiment that the interaction of marine biology with the physical system may influence some of the observed interannual and decadal variability in the Pacific, but a detailed exploration of such complex behavior will require longer timeseries and additional sensitivity studies.

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References

- Alexander, M., I. Blade, M. Newman, J. Lanzante, N.-C. Lau, and J. Scott, 2002: The atmospheric bridge: The influence of ENSO teleconnections on air-sea interactions over the global oceans. *J. Climate*, **15**, 2205–2231.
- Chavez, F., P. Strutton, and M. McPhaden, 1998: Biological-physical coupling in the central equatorial Pacific during the onset of the 1997-98 El Niño. *Geophys. Res. Lett.*, **25**, 3543–3546.
- Cloern, J., C. Grenz, and L. Videgar-Lucas, 1995: An empirical model of the phytoplankton:carbon ratio – the conversion factor between productivity and growth rate. *Limnology and Oceanography*, **40**, 1313.

- Federov, A. V. and S. G. Philander, 2001: A stability analysis of tropical ocean-atmosphere interactions: Bridging measurements and theory for El niño. *J. Climate*, **14**, 3086–3101.
- Fouquart, Y. and B. Bonnel, 1980: Computations of solar heating of the earth's atmosphere: A new parameterization. *Beitr. Phys. Atmos.*, **53**, 35–62.
- Gent, P., J. Willebrand, T. McDougall, and J. McWilliams, 1995: Parameterizing eddy-induced tracer transport in ocean circulation models. *J. Phys. Oceanogr.*, **25**, 463–474.
- Heinze, C., A. Hupe, E. Maier-Reimer, N. Dittert, and O. Ragueneau, 2003: Sensitivity of the marine biospheric Si cycle for biogeochemical parameter variations. *Global Biogeochem. Cycles*, **17**, doi:10.1029/2002GB001943.
- Hibler, W., 1979: A dynamic thermodynamic sea ice model. *J. Phys. Oceanogr.*, **9**, 815–846.
- Jungclaus, J., M. Botzet, H. Haak, N. Keenlyside, J.-J. Luo, M. Latif, J. Marotzke, U. Mikolajewicz, and E. Roeckner, 2004: Ocean circulation and tropical variability in the AOGCM ECHAM5/MPI-OM. *J. Climate*.
- Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Woolen, Y. Zhu, M. Chelliah, W. Ebisuzaki, W. Higg, A. Leetsmaa, R. Reynolds, R. Jenne, and D. Joseph, 1996: The NCEP/NCAR 40 year-reanalysis project. *Bull. Amer. Meteor. Soc.*, **77**, 437–470.
- Kara, A., H. HURLBURT, P. ROCHFORD, and J. O. BRIEN, 2004: The impact of water turbidity on interannual sea surface temperature simulations in a layered global ocean model. *jpo*, **34**, 345–359.
- Lin, S. J. and R. B. Rood, 1996: Multidimensional flux form semi-lagrangian transport. *Mon. Wea. Rev.*, **124**, 2046–2068.
- Maier-Reimer, E., 1993: Geochemical cycles in an ocean general circulation model. preindustrial tracer distribution. *Global Biogeochem. Cycles*, **7**, 645–677.
- Manizza, M., C. L. Quéré, A. Watson, and E. Buitenhuis, 2004: Bio-optical feedbacks among phytoplankton, upper ocean physics and sea-ice in a global model. *submitted to GRL*.
- Marsland, S. J., H. Haak, J. H. Jungclaus, M. Latif, and F. Roeske, 2003: The Max-Planck-Institute global ocean/sea ice model with orthogonal curvilinear coordinates. *Ocean Modelling*, **5**, 91–127.
- Marzeion, B., A. Timmermann, R. Murtugudde, and F.-F. Jin, 2004: Bio-physical feedbacks in the tropical Pacific. *J. Climate*, **in press**.
- Miller, A. e. a., 2003: Potential feedbacks between pacific ocean ecosystems and interdecadal climate variations. *Bull. Amer. Meteor. Soc.*, **84**, 617–633.
- Moore, J., S.C.Doney, J. Kleypas, D. Glover, and I.Y.Fung, 2002: An intermediate complexity marine ecosystem model for the global domain. *Deep-Sea Research II*, **49**, 403–462.
- Morel, A. and S. Maritorena, 1988: Optical modeling of the upper ocean in relation to its biogenous matter content (case 1 waters). *J. Geophys. Res.*, **93**, 10749–10768.
- 2001: Bio-optical properties of oceanic waters: A reappraisal. *J. Geophys. Res.*, **106**, 7163–7180.
- Murtugudde, R., J. Beauchamp, C. McClain, M. Lewis, and A. Busalacchi, 2002: Effects of penetrative radiation on the upper tropical ocean circulation. *J. Climate*, **15**, 470–486.
- Nakamoto, S., S. Kumar, J. Oberhuber, K. Muneyama, and R. Frouin, 2000: Chlorophyll modulation of sea surface temperature in the arabian sea in a mixed-layer isopycnal general circulation model. *Geophys. Res. Lett.*, **27**, 747–750.
- Nakamoto, S., S. P. Kumar, J. Oberhuber, K. Muneyama, and R. Frouin, 2001: Response of the equatorial pacific to chlorophyll pigments in a mixed-layer isopycnal general circulation model. *Geophys. Res. Lett.*, **28**, 2021–2024.

- Ohlmann, J. and D. Siegel, 2000: Ocean radiant heating. part ii: Parameterizing solar radiation transmission through the upper ocean. *J. Phys. Oceanogr.*, **30**, 1849–1865.
- Ohlmann, J., D. Siegel, and C. Mobley, 2000: Ocean radiant heating. part i: Optical influences. *J. Phys. Oceanogr.*, **30**, 1833–1848.
- Rayner, N., D. Parker, E. Horton, C. Folland, L. Alexander, D. Rowella, E. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**, doi:10.1029/2002JD002670.
- Roeckner, E., G. Baeuml, L. Bonventura, R. Brokopf, M. Esch, M. Giorgetta, S. Hagemann, I. Kirchner, L. Kornblueh, E. Manzini, A. Rhodin, U. Schlese, U. Schulzweida, and A. Tompkins, 2003: The atmospheric general circulation model echam 5. part i: Model description. Report 349, Max Planck Institute for Meteorology, Hamburg, Germany, available from <http://www.mpimet.mpg.de>.
- Shell, K. M., R. Frouin, S. Nakamoto, and R. Somerville, 2003: Atmospheric response to solar radiation absorbed by phytoplankton. *JGR*, **108**, doi:10.1029/2003JD003440.
- Siegel, D., J. C. Ohlmann, L. Washburn, R. Bidigare, C. Nosse, E. Fields, and Y. Zhou, 1995: Solar radiation, phytoplankton pigments and radiant heating of the equatorial pacific warm pool. *J. Geophys. Res.*, **100**, 4885–4891.
- Six, K. and E. Maier-Reimer, 1996: Effects of plankton dynamics on seasonal carbon fluxes in an ocean general circulation model. *Global Biogeochem. Cycles*, **10**, 559–583.
- Strutton, P. and F. Chavez, 2003: Biological heating in the equatorial pacific: Observed variability and potential for real-time calculation. *J. Climate*, **17**, 1097–1109.
- Sweby, P., 1984: High resolution schemes using flux limiters for hyperbolic conservation laws. *SIAM J. Numer. Anal.*, **21**, 995–1011.
- Timmermann, A. and F.-F. Jin, 2002: Phytoplankton influences on tropical climate. *Geophys. Res. Lett.*, **29**, doi:10.1029/2002GL0015434,2002.
- Timmreck, C. and M. Schulz, 2004: Significant dust simulation differences in nudged and climatological operation mode of the AGCM ECHAM. *J. Geophys. Res.*, **109**, doi:10.1029/2003JD004381.
- Valcke, S., A. Caubel, D. Declat, and L. Terray, 2003: Oasis ocean atmosphere sea ice soil users's guide. Technical Report TR/CMGC/03/69, CERFACS, Toulouse, France.