Atmospheric response to solar radiation absorbed by phytoplankton

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[1] Phytoplankton alter the absorption of solar radiation, affecting upper ocean temperature and circulation. These changes, in turn, influence the atmosphere through modification of the sea surface temperature (SST). To investigate the effects of the present-day phytoplankton concentration on the atmosphere, an atmospheric general circulation model was forced by SST changes due to phytoplankton. The modified SST was obtained from ocean general circulation model runs with space- and time-varying phytoplankton abundances from Coastal Zone Color Scanner data. The atmospheric simulations indicate that phytoplankton amplify the seasonal cycle of the lowest atmospheric layer temperature. This amplification has an average magnitude of 0.3°K but may reach over 1°K locally. The surface warming in the summer is marginally larger than the cooling in the winter, so that on average annually and globally, phytoplankton warm the lowest layer by about 0.05°K. Over the ocean the surface air temperature changes closely follow the SST changes. Significant, often amplified, temperature changes also occur over land. The climatic effect of phytoplankton extends throughout the troposphere, especially in middle latitudes where increased subsidence during summer traps heat. The amplification of the seasonal cycle of air temperature strengthens tropical convection in the summer hemisphere. In the eastern tropical Pacific Ocean a decreased SST strengthens the Walker circulation and weakens the Hadley circulation. These significant atmospheric changes indicate that the radiative effects of phytoplankton should not be overlooked in studies of climate change. *INDEX TERMS:* 0315 Atmospheric Composition and Structure: Biosphere/atmosphere interactions; 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 4855 Oceanography: Biological and Chemical: Plankton; 4847 Oceanography: Biological and Chemical: Optics; *KEYWORDS:* phytoplankton, atmospheric general circulation model (AGCM), absorption of solar radiation, seasonal cycle, sea surface temperature (SST)

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1. Introduction

[2] Phytoplankton and their covarying materials dominate the variability of the optical properties of seawater in the open ocean. Phytoplankton absorb solar radiation, especially in the 350 to 700 nm spectral range (about half of the solar flux), thus decreasing the solar heating below the mixed layer [*Lewis et al.*, 1990; *Siegel et al.*, 1995; *Ohlmann et al.*, 2000]. The effect of this absorption on ocean temperature is dependent on the relative depths of radiation attenuation and the mixed layer. If the mixed layer is shallow, it is particularly sensitive to changes in phytoplankton. Deep mixed layers are less sensitive to changes in phytoplankton since the amount of penetrating solar radiation is small regardless of whether phytoplankton are present or not. Thus phytoplankton are especially important in shallow mixed layer regions, where they tend to heat the mixed layer, increasing sea surface temperature (SST) [*Ohlmann et al.*, 1996; *Sathyendranath et al.*, 1991]. These heating changes may cause dynamical changes in the ocean mixed layer depth and circulation, which themselves can influence the temperature profile [*Lewis et al.*, 1983; *Ohlmann et al.*, 1998].

[3] Ocean general circulation models (OGCMs) often assume that all the solar irradiance penetrating the surface is absorbed in the mixed layer or in the top model layer. The total absorption in models was identified as a problem by *Lewis et al.* [1990], who showed that in the tropical Pacific

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Ocean a significant amount of sunlight at visible wavelengths effectively penetrates below the mixed layer. These authors argued that ocean transparency in this region may explain SST overestimation by OGCMs that neglect sunlight penetration. The results of *Lewis et al.* [1990] were confirmed in coupled atmosphere-ocean GCM experiments performed by *Schneider and Zhu* [1998], who improved the simulation of the warm pool in the western Pacific Ocean by specifying a global attenuation depth for sunlight penetration.

[4] *Schneider and Zhu* [1998] used the same attenuation depth for the entire ocean, ignoring the spatial and seasonal variation of phytoplankton. Recently, the variable absorption of solar radiation by phytoplankton has been incorporated into OGCMs using the space and time distribution of phytoplankton concentration from satellite ocean color data. *Nakamoto et al.* [2000, 2001] compared SST results from runs with the spatial and seasonal pattern of absorption to those with a constant attenuation depth of 23 m (corresponding to the clearest waters) to study the net effect of phytoplankton in the Arabian Sea and equatorial Pacific. *Murtugudde et al.* [2002] determined the spatial pattern of attenuation depth in a tropical ocean model and compared the resulting SST field to that obtained using a constant attenuation depth of 17 m, the global mean value. The two studies used different control attenuation depths and thus obtained different SST results. However, in both cases, the heating changes due to phytoplankton resulted in dynamical changes, such as altered currents and upwelling, which changed the magnitude and sometimes the sign of the SST difference.

[5] While many studies have explored ocean responses to phytoplankton, the atmospheric response to SST perturbations related to phytoplankton has not been previously investigated. The purpose of this study is to make a first attempt at understanding how the present phytoplankton distribution influences the atmosphere indirectly through altering SST. Our method is first to use an OGCM with and without the radiative effects of phytoplankton to determine how phytoplankton change SST, as described in section 2. We then introduce the SST changes from the ocean model as boundary conditions for an AGCM, presented in section 3. The results are discussed in section 4 and provide an estimate of how different the atmospheric temperature and circulation would be if there were no phytoplankton (i.e., a "dead" ocean). In section 5, we summarize the results and draw conclusions about the importance of radiative forcing by phytoplankton in studies of climate change.

[6] There is, of course, feedback within the ocean, atmosphere, and phytoplankton system, which is ignored when ocean and atmospheric models are run separately. In this study, we have chosen to explore how imposed SST changes due to phytoplankton may affect the large-scale atmospheric circulation. If the impact is significant in this uncoupled system, then the radiative effects of phytoplankton could play an important role in coupled models and climate change simulations, and further work to study the coupled system and its feedbacks will be justified.

2. Effect of Phytoplankton on SST

[7] We use an updated version of the Max Plank Institute (MPI) ocean isopycnal model (OPYC), developed by *Oberhuber* [1993], to examine the effects of phytoplankton on SST. The results are those used by *Nakamoto et al.* [2000, 2001]. The OPYC model includes a realistic equation of state, solves the primitive equations, and fully couples surface mixed layer, snow, and sea ice models to the ocean interior. The domain is the world's oceans with 322×152 horizontal grid cells and twelve vertical isopycnal layers below a turbulent surface mixed layer. Monthly mean atmospheric data from the European Center for Medium-Range Weather Forecast (ECMWF) reanalysis, interpolated to daily values, are used for surface forcing.

[8] Two different parameterizations of ultraviolet and visible solar heating were used. (Infrared fluxes are treated separately.) The first one, based on *Paulson and Simpson* [1977], assumes a 23 m solar radiation attenuation length everywhere, approximating the lowest phytoplankton concentrations. The second parameterization, by *Morel and Antoine* [1994], computes the vertical profile of the heating rate on the basis of surface phytoplankton pigment concentration as measured from satellite ocean color sensors. The parameterization includes the effects of phytoplankton and their covarying derivative products (e.g., detritus and dissolved organic matter). A monthly mean climatology of pigment concentration from Coastal Zone Color Scanner (CZCS) data [*Feldman et al.*, 1989] was used for this second parameterization. The CZCS climatology describes the annual cycles of phytoplankton biomass well, in agreement with conceptual and mathematical models of plankton dynamics [*Yoder et al.*, 1993]. The pigment concentration is assumed to be constant with depth. This assumption is acceptable since at depth the residual radiant energy is low [see *Morel and Antoine*, 1994].

[9] The OPYC model was run for 50 years using Paulson and Simpson's parameterization to obtain a cyclo-stationary state. At the end of this run, the interannual variation of the global mean SST was less than 0.05°K. The final state of the 50th year of this integration was used as the initial condition for the chlorophyll-forcing experiment [*Nakamoto et al.*, 2000, 2001].

[10] The model was run for an additional 10 years with Paulson and Simpson's parameterization (control run) and 10 years with Morel and Antoine's parameterization (phytoplankton run), starting from the same initial conditions. At the end of these runs, new cyclo-stationary states were obtained, with interannual variations of global mean SST less than 0.05°K between the 59th and 60th years. The SST fields from the last (60th) years for each run are used in the analysis. Results in the southern oceans poleward of 60°S may not be significant because of uncertainties in the satellite-derived pigment concentration.

[11] The SST difference between the phytoplankton run and the control run (Figure 1) represents the net effect of the present phytoplankton concentration on SST, since the default parameterization [i.e., *Paulson and Simpson*, 1977] corresponds closely to the clearest oceanic waters. In the summer hemisphere, the phytoplankton run generally has a higher SST than the control run (Figures 1a and 1b), because more solar radiation is absorbed in the mixed layer and less penetrates to lower layers. In the fall and early winter, the ocean surface is warmer and thus able to transfer heat to the atmosphere more efficiently through increased infrared emission and latent and sensible heat fluxes. Therefore the ocean loses more heat in the phytoplankton run. When the

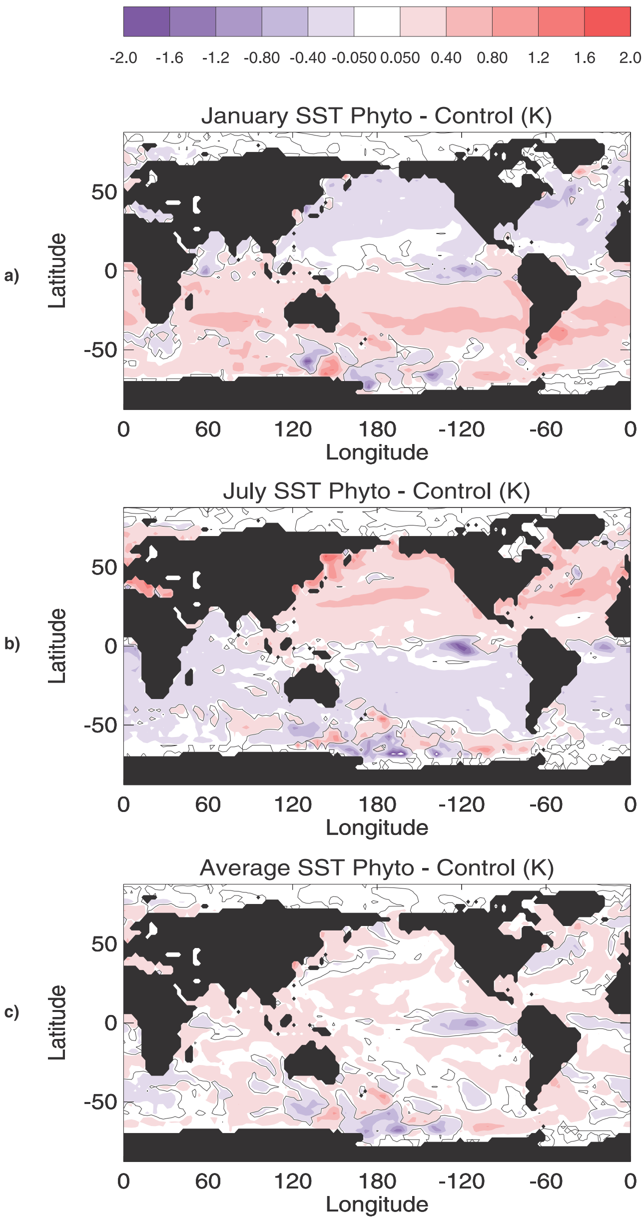

Figure 1. SST difference between the phytoplankton and control ocean model runs for (a) January, (b) July, and (c) the annual average. The line indicates the zero contour.

mixed layer deepens in the winter, it brings up water from below which is colder than the corresponding water in the control run, resulting in a cooler winter mixed layer. Thus, in the winter hemisphere, the average SST is cooler in the phytoplankton run than in the control run. The annual effect of phytoplankton is generally positive, but of a smaller magnitude than individual monthly effects (Figure 1c).

[12] Therefore the major effect of including phytoplankton in an ocean model is an amplification of the seasonal cycle. The seasonally varying phytoplankton concentration increases the amplitude of the annual cycle of SST by about 0.3°K in both the Northern Hemisphere and Southern Hemisphere, as shown in Figure 2. This amplification corresponds to about 20% of the seasonal cycle of SST in most oceans, except in equatorial regions where local buoyancy plays a lesser role in the dynamics of the mixed layer. Over the year, the warming in the summer is slightly larger than the cooling in the winter; thus phytoplankton increase the annually averaged SST by about 0.04°K.

[13] The other significant effect of phytoplankton is the persistent SST decrease in the equatorial Pacific around 110°W. *Nakamoto et al.* [2001] hypothesized that mixed layer heating in this region due to the high phytoplankton

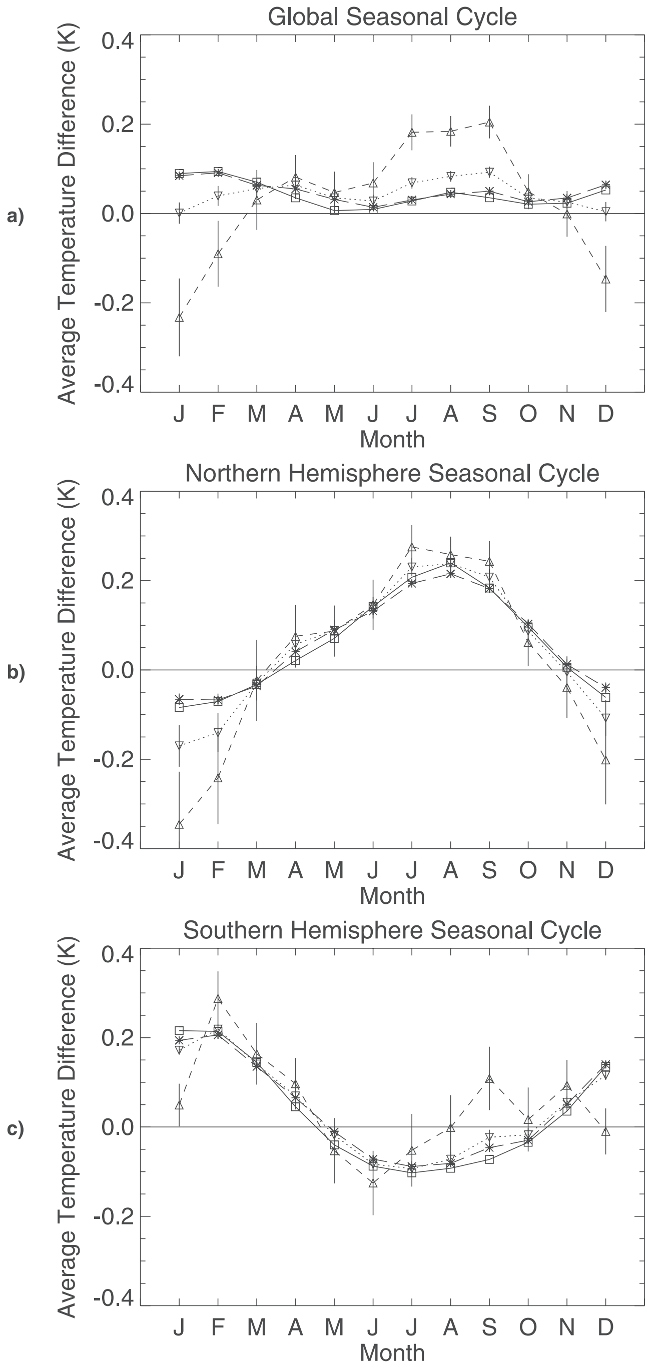

Figure 2. Monthly average (a) global, (b) Northern Hemisphere, and (c) Southern Hemisphere temperature differences (phytoplankton run-control run) between 60°S and 60°N for SST (squares) and lowest-layer atmospheric temperature of all points (upside-down triangles), ocean points (asterisks), and land points (triangles). The vertical lines indicate the standard error of the difference.

concentration decreases mixed layer depth at the equator, generating anomalous geostrophic westward currents north and south of the equator and consequently strengthening the eastward equatorial undercurrent. An increased equatorial undercurrent enhances upwelling in the east because of mass convergence, resulting in a decreased SST in the phytoplankton run. This process may also occur during some months in the tropical Atlantic (e.g., in July; see Figure 1b).

[14] Since the OGCM was run with the climatological atmosphere and phytoplankton, these results do not include any feedbacks between the ocean temperature, mixed layer depth, or circulation and the atmosphere or phytoplankton. However, these ocean model experiments provide a first-order estimate of the effect of the observed phytoplankton on SST. On the basis of these results, we expect an ocean without phytoplankton (i.e., a "dead" ocean) to have a weaker seasonal cycle compared to the present-day.

3. Atmospheric Model Simulations

[15] To determine the effects, if any, of absorption due to phytoplankton on the atmosphere, we used the NCAR Community Climate Model version 3 (CCM3) [*Kiehl et al.*, 1998], an AGCM. CCM3 is a spectral model with T42 resolution (128 longitudinal $\times$ 64 latitudinal values), 18 vertical levels, and a time step of 20 minutes. The model is coupled to a land surface model [*Bonan*, 1998]. CCM3 includes parameterizations of clouds, radiation, boundary layer processes, gravity wave drag, Rayleigh friction, convection, and stable condensation.

[16] For the control runs, we used yearly cycling climatological SSTs as a boundary condition (12 monthly values, with interpolation between the monthly values). We created a yearly cycling SST set for the plankton runs by summing the climatological SSTs and the monthly average SST differences due to phytoplankton from the ocean model. Since CCM3 identifies grid points with sea ice as those having an SST below $-1.8°C$, we adjusted the phytoplankton SST field so that sea ice did not change locations between the control and phytoplankton run.

[17] We performed two control runs and three plankton runs. The first control run and the first plankton run were made in parallel for 18 years, starting from the same initial condition. In addition, the first plankton run continued for another 9 years. The second plankton run started with the state of the system one year into the first control run and ran for 10 years. The third plankton run started with the state two years into the first control run and ran for 12 years. The second control run started one year into the first phytoplankton run and ran for 30 years. The first year of each run was not used for data analysis, in order to allow the atmosphere to adjust to the imposed SST field. Thus there were 46 usable control years and 46 usable plankton years. This length of time is sufficient to achieve a robust climatological average, except in the high latitudes, where the variance is quite high.

[18] We examined monthly averages of model variables, concentrating on temperature in the lowest atmospheric layer and on convective precipitation. Convective precipitation is the sum of precipitation calculated by the *Zhang and McFarlane* [1995] deep convection scheme and the *Hack* [1994] shallow/midlevel convection scheme. It excludes precipitation associated with large-scale ascent. The monthly values from all the phytoplankton runs were averaged to obtain annual means and monthly means for each month of the year. The monthly values for the control run were similarly averaged. The differences between the means (plankton minus control) provide a measure of the effect of the phytoplankton. The significance of the differences was determined using a Student's *t* test.

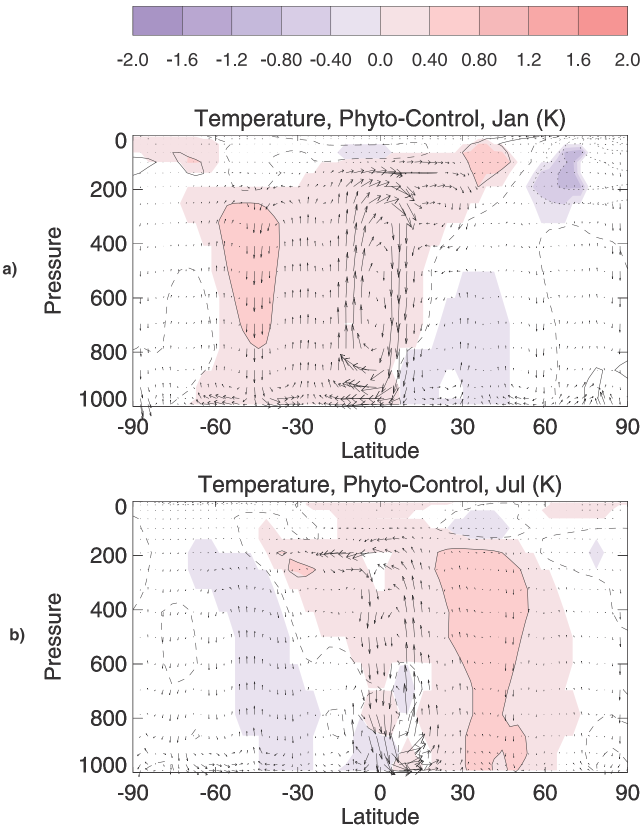

Figure 3. Difference between phytoplankton and control runs for (a) January and (b) July longitudinally averaged temperature (colored contours) and circulation (arrows). Solid contours correspond to positive temperature differences, while dotted contours indicate negative temperature differences. The dashed line follows the zero contour. Temperature differences with a significance of at least 95% are shaded.

4. Results and Discussion

[19] The primary effect of the phytoplankton SST is an amplification of the seasonal cycle in the lowest atmospheric layer temperature (Figure 2) by about 0.3 K, similar to the SST seasonal cycle amplification. The amplification of the seasonal cycle is clearest over the ocean. Because the lowest atmospheric layer temperature is strongly influenced by the underlying SST, this seasonal cycle amplification is very similar between runs and closely matches the cycle amplification found in the SST field. The increase in the temperature cycle over land is not as smooth as that of the SST cycle because of intensified air temperature changes over land. The amplification of the seasonal cycle extends throughout the troposphere (Figure 3).

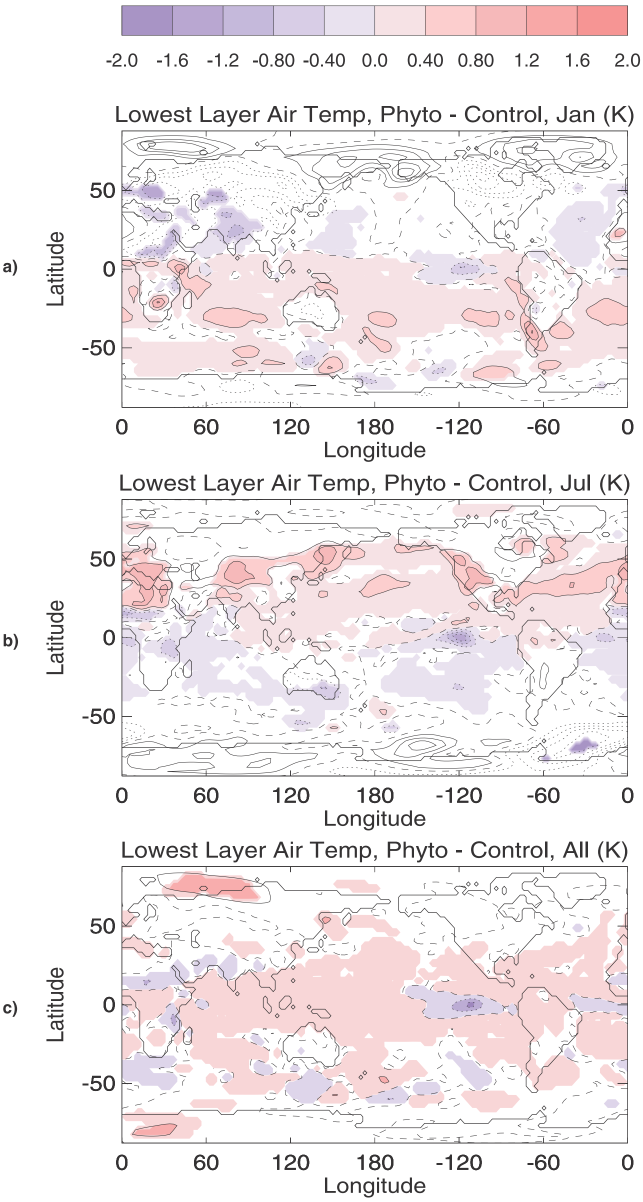

Figure 4. (a) January, (b) July, and (c) yearly average lowest-layer temperature difference between phytoplankton and control runs. Solid contours correspond to regions that are warmer with phytoplankton, while dotted contours indicate cooler regions. The dashed line follows the zero contour. Differences with a significance of at least 95% are shaded.

[20] In the annual average, the phytoplankton run is slightly warmer overall than the control run, by about 0.05°K. Over the ocean, the annual temperature changes in the lowest atmospheric layer closely follow the imposed SST differences (Figure 4c). Most of the ocean point differences are positive; however, the eastern tropical Pacific and equatorial Atlantic are persistently cooler, corresponding to the similar decreases in SST. Land temperatures show a larger variation in response since they are not directly forced.

[21] Annual convective precipitation changes are found in the tropics (Figure 5c). Convective precipitation decreases in the eastern equatorial Pacific and increases north of the eastern equatorial Pacific and in the western equatorial Pacific, by about 1 mm/day. These changes are related to the persistent negative SST forcing in the eastern tropical Pacific. Annual mean changes are also found in the western Indian Ocean and tropical Atlantic, which receive significantly less precipitation in the presence of phytoplankton. The convective changes are confined to tropical regions; no large-scale changes to precipitation are found in the middle latitudes.

[22] The seasonal responses in the tropics and middle latitudes are further examined in sections 4.1 and 4.2.

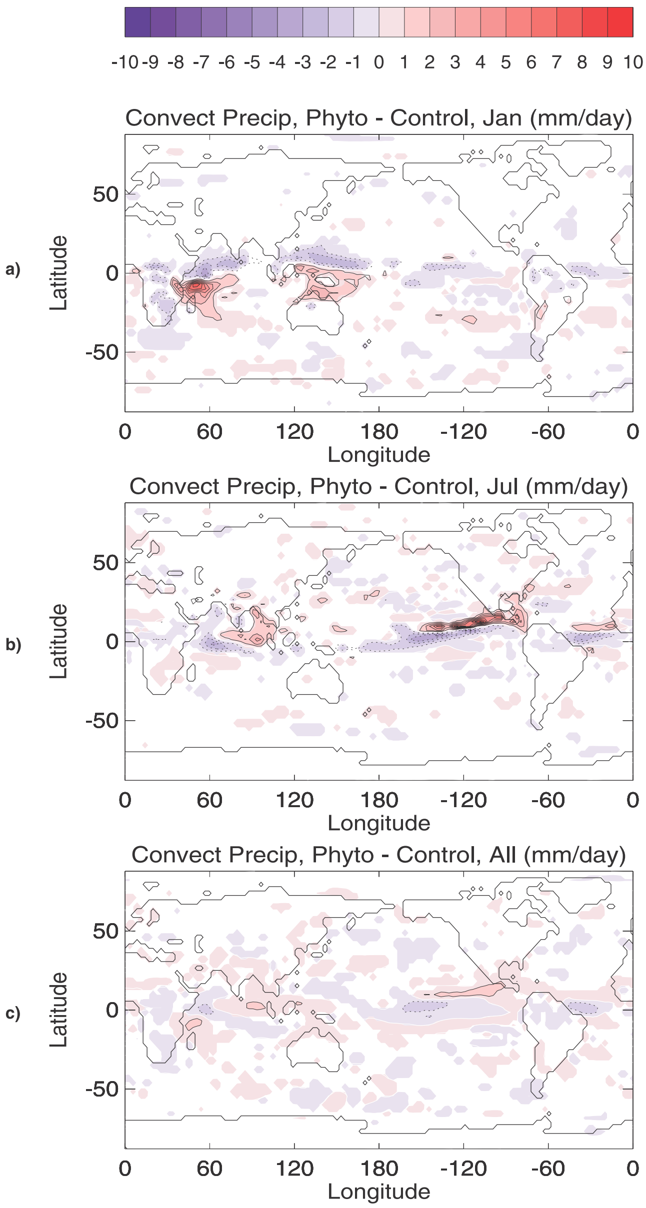

Figure 5. (a) January, (b) July, and (c) yearly average convective precipitation differences between phytoplankton and control runs. Solid contours correspond to additional precipitation, while dotted contours indicate less precipitation. The zero contour is not included. Differences with a significance of at least 95% are shaded.

4.1. Tropics

[23] In the tropics, the changes in convective precipitation are well correlated to the specified SST differences. For example, in January, the northern part of the western Indian ocean is colder while the southern part is warmer. Thus convective precipitation decreases in the north and increases in the south. These precipitation changes correspond to specific humidity changes (not shown). A similar mechanism occurs in the western tropical Pacific, causing a dipole pattern with less precipitation to the north and more precipitation to the south, even though the changes in SST are very small (less than 0.1°K). Likewise, in July, the western Indian Ocean is cooler and drier, and the eastern oceans show increased convection to the north and decreased convection to the south.

[24] Overall, these changes are associated with an increase of convection and precipitation in the summer hemisphere in response to the amplification of the seasonal cycle in the phytoplankton run. The ITCZ is not significantly displaced; the precipitation changes occur locally. The enhanced precipitation in the summer hemisphere corresponds to a stronger upward flow around 5°S in January and 5°N in July (Figure 3).

[25] The persistent decreased SST in the eastern tropical Pacific results in circulation changes in the tropics (Figures 6 and 7). The decrease in convection in this region weakens the Hadley circulation locally, decreasing the meridional convergence near the surface. (The strengths of the northerly winds in the northern part of the region and the southerly winds in the southern part are decreased.) At the surface north and south of the decreased SST, the surface easterlies weaken because of the Coriolis force on the surface divergence. In addition, this relatively cold region increases the zonal tropical SST gradient, which strengthens the Walker circulation, resulting in a stronger easterly wind at the equator. When this cold SST region is present in the Atlantic Ocean (e.g., Figure 1b), a similar response occurs.

4.2. Middle and High Latitudes

[26] Middle latitude atmospheric temperature follows the imposed SST changes, with significant temperature increases in summer and decreases in winter (Figures 4a and 4b). Effects over land generally reflect the nearby SST forcing. For example, in July, phytoplankton increase the Mediterranean Sea SST, which warms the surrounding region. Likewise, warmer July temperature in eastern Asia and western North America reflect the large coastal SST changes of the Pacific Ocean. In January, a colder Mediterranean Sea and north Indian Ocean cool the nearby land points. The temperature effects over land regions adjacent to the ocean are correlated with the average prevailing wind direction at the Earth's surface.

[27] The strong temperature changes found in winter high latitudes are not statistically significant. Because of the high model variability in these regions, more runs would be needed to achieve a statistically steady state. Note, however, that enhanced subsidence around 50°S in January and 50°N in July tends to trap heat in the troposphere (Figure 3). This effect is amplified in the Northern Hemisphere by interactions between the land surface and the atmosphere.

[28] Finally, in January in the Southern Hemisphere, the easterlies and westerlies associated with anticyclonic gyres of the Indian and Pacific Oceans are altered (Figure 6a),

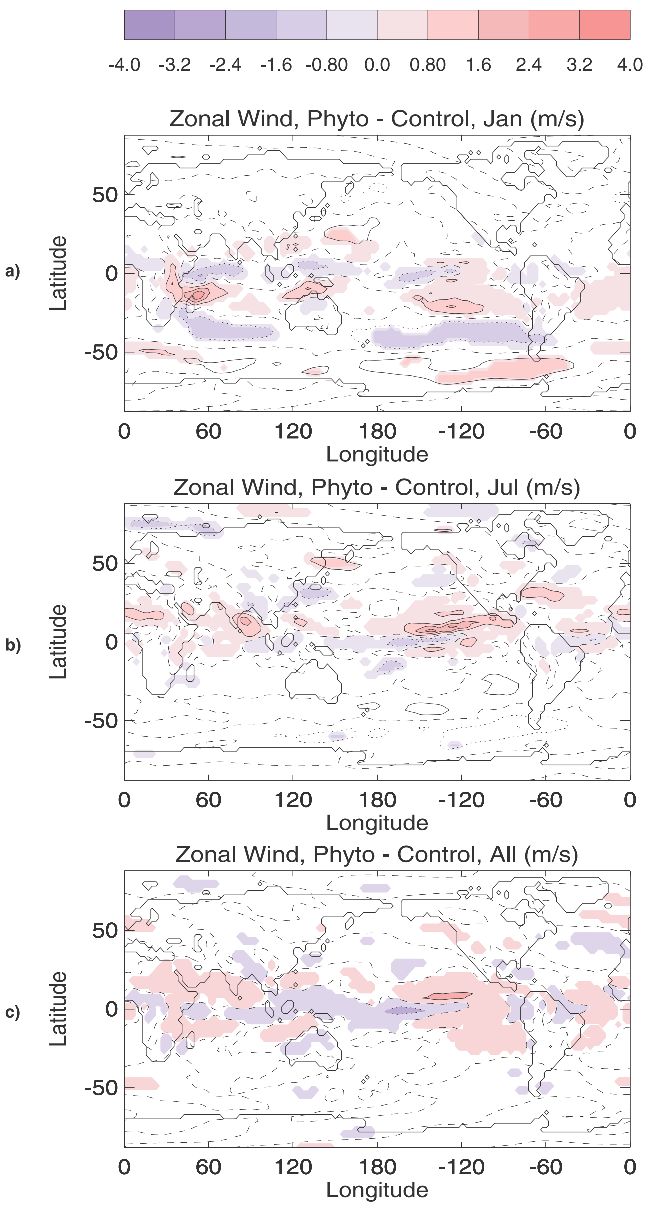

Figure 6. (a) January, (b) July, and (c) yearly average lowest atmospheric layer zonal wind changes. The shaded regions indicate points where the magnitude of the difference has a significance of at least 95%. (Because of a data storage problem, a few years of data are missing from these figures; however, the basic pattern of tropical wind changes are present in all the remaining years, so the lack of data probably does not significantly affect the results.)

consistent with a weakening of the high pressure toward the equator and a strengthening poleward, as seen in the surface pressure field (not shown). In July in the Northern Hemisphere (Figure 6b), a similar, yet much less coherent pattern of change is obtained.

5. Conclusions

[29] The primary atmospheric effect of phytoplankton is the amplification of the seasonal cycle of lowest-layer atmospheric temperature. The increases in temperature

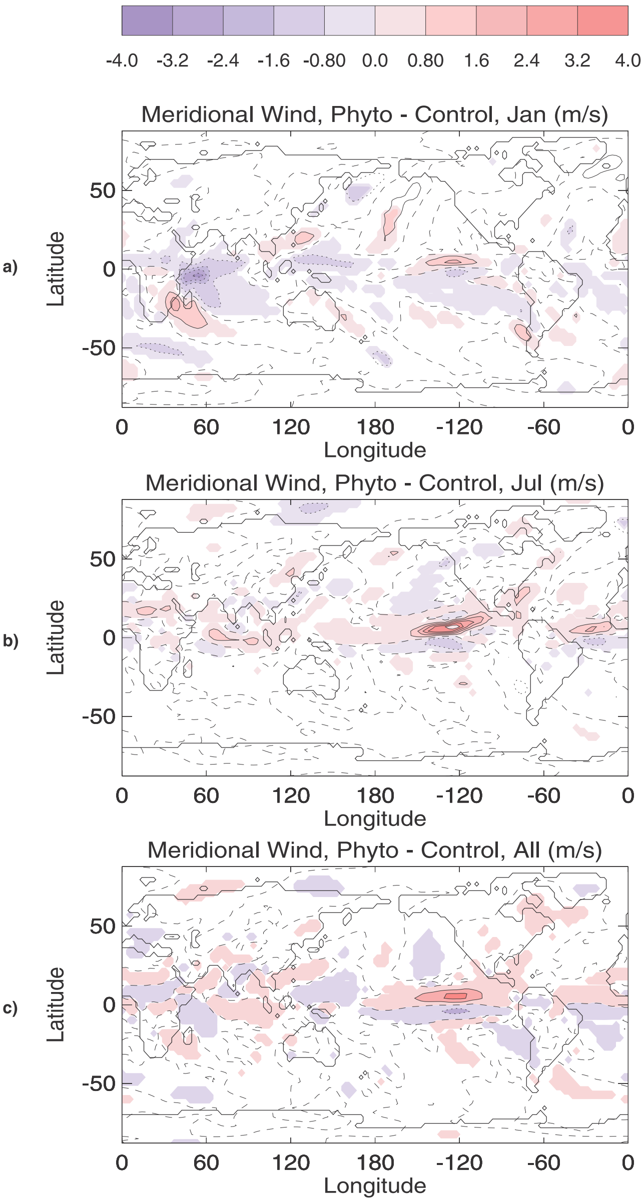

Figure 7. (a) January, (b) July, and (c) yearly average lowest atmospheric layer meridional wind changes. The shaded regions indicate points where the magnitude of the difference has a significance of at least 95%. (Because of a data storage problem, a few years of data are missing from these figures; however, the basic pattern of tropical wind changes are present in all the remaining years, so the lack of data probably does not significantly affect the results.)

during the summer are slightly larger than the decreases in the winter, so the annually averaged effect is one of slight warming. In response to the increased summer temperature, convection intensifies in the summer hemisphere. Changing patterns of tropical convection lead to atmospheric circulation changes. In addition, the persistent negative SST difference in the eastern tropical Pacific strengthens the Walker circulation and weakens the Hadley circulation in this region.

[30] The midlatitudes do not show significant convective changes; however, we do find temperature changes over land of up to 1°K. These changes are often higher in magnitude than those found over the ocean. While temperature differences over the ocean are constrained by the specified SST forcing, land point changes are not directly regulated by the ocean and can thus vary over a wider range.

[31] By separating the ocean and atmosphere models, this work neglected important feedbacks within the system, which could alter the effect of the phytoplankton. The atmosphere's response to an SST change generally reduces the magnitude of the change as heat is transferred between the ocean and the atmosphere. In an atmospheric model experiment with a fixed SST, such as this work, the ocean is essentially an infinite heat source. Thus we expect that our results overestimate the magnitude of the effect of phytoplankton. However, local positive feedbacks may amplify the original effect in some regions. To more fully examine the effects, a coupled ocean-atmosphere model is required. However, the major effects, the amplified seasonal cycle and changes in tropical convection, are quite robust and probably would be obtained in a coupled model experiment.

[32] These model runs, by specifying the phytoplankton concentration, also excluded possible feedbacks between phytoplankton, the surrounding ocean, and the atmosphere. The amplified seasonal cycle of SST may reduce mixing rates during the summer and therefore nutrients in the euphotic zone, reducing the phytoplankton biomass in summer. In addition, changes in solar irradiance, precipitation, and circulation may affect light and nutrient availability for photosynthesis and phytoplankton growth. For example, enhanced precipitation in the summer hemisphere stabilizes the mixed layer, reducing the supply of nutrients from below and decreasing the amount of phytoplankton that can be supported in the region. Furthermore, enhanced precipitation over Northern Australia and Indonesia in January and over Southeast Asia and the Bay of Bengual in July may reduce biomass burning aerosols in those tropical regions, with consequences on local radiative forcing and climate. The wind changes may also affect the phytoplankton distribution through changes in nutrient sources. Over the African continent north of the ITCZ, the easterly winds are weakened in July (Figure 6b), possibly reducing the aeolian supply of dust (i.e., iron) to the tropical Atlantic. Off the coast of East Africa south of the equator, enhanced northerly winds in January (Figure 7a) increase upwelling of nutrients, but the effect is opposed by enhanced precipitation (Figure 5a). Thus the present-day concentration may be influenced through its interaction with the rest of the climate system.

[33] This work indicates that numerical ocean and coupled models should include the space and time distribution of the absorption of visible solar radiation by phytoplankton. (Note that atmospheric models that use the observed SSTs already include the space- and time-dependent radiative effects of phytoplankton.) Most OGCMs treat solar absorption the same way for the entire ocean, independent of location or time, by either specifying a global attenuation depth or absorbing all solar radiation within the mixed layer. Neglecting the variance of phytoplankton may significantly affect the heat capacity and upwelling of the oceans as well as tropical convection patterns and atmospheric circulation. Adding the time-space variance of solar penetration may not

immediately improve ocean models, since many models have been tuned to approximate the observations. However, it is important to understand all the processes which can significantly affect ocean model results.

[34] Our results emphasize that phytoplankton could play an important role in climate and potential climate change through their interaction with radiation. In order to investigate effects on future climate, one needs to know how phytoplankton abundance and type (i.e., optical properties) will respond to changing conditions. Growth rates and metabolic processes, hence biological production, would be directly affected by temperature and sunlight changes. Phytoplankton concentrations would also be affected by changes in nutrient availability. Attempts have been made to predict in situ and satellite observations of chlorophyll concentration using three-dimensional coupled biological/physical models on basin and global scales [e.g., *Gregg and Walsh*, 1992; *Sarmiento et al.*, 1993; *Walsh et al.*, 1999; *Gregg*, 2002]. Although some reasonable predictions have resulted, limited success has been achieved because of uncertainties in the mechanistic equations governing the evolution of prognostic variables. It might be possible, however, to predict phytoplankton changes from state variables such as sunlight, temperature, and biomass. A statistical approach by biological province, such as those classified by *Longhurst et al.* [1995], using available long-term satellite data sets of ocean color, SST, and solar irradiance may be envisioned. Further research in this area may provide, for studies of climate change incorporating phytoplankton radiative forcing, a suitable alternative to fully coupled biological/physical models.

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