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Quantification of the Feedback between Phytoplankton and ENSO in the Community Climate System Model

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ABSTRACT

The current coarse-resolution version of the Community Climate System Model is used to assess the impact of phytoplankton on El Niño–Southern Oscillation (ENSO). The experimental setup allows for the separation of the effects of climatological annual cycle of chlorophyll distribution from its interannually varying part. The main finding is that the chlorophyll production by phytoplankton is important beyond modifying the mean and seasonal cycle of shortwave absorption; interannual modifications to the absorption have an impact as well, and they dampen ENSO variability by 9%. The magnitude of damping is the same in the experiment with smaller-than-observed, and in the experiment with larger-than-observed, chlorophyll distribution. This result suggests that to accurately represent ENSO in GCMs, it is not sufficient to use a prescribed chlorophyll climatology. Instead, some form of an ecosystem model will be necessary to capture the effects of phytoplankton coupling and feedback.

1. Introduction

Phytoplankton and chlorophyll affect the absorption of shortwave radiation and thus the vertical distribution of heat in the upper ocean (Lewis et al. 1990). In the tropical Pacific in particular, this interaction can lead to a feedback between atmospheric circulation and chlorophyll because of the combination of strong radiation, large equatorial chlorophyll concentrations (Shell et al. 2003), and the fact that the tropical atmosphere is very sensitive to sea surface temperature (SST) changes (Palmer and Mansfield 1984).

Ocean general circulation models (OGCMs), too, have shown a connection between chlorophyll concentrations and equatorial SST (Nakamoto et al. 2001; Murtugudde et al. 2002; Manizza et al. 2005; Sweeney et al. 2005; Löptien et al. 2009), and so have coupled ocean–atmosphere general circulation models (GCMs). In response to increased chlorophyll concentrations, a reduction in ENSO amplitude is reported in one full GCM (Wetzel et al. 2006), and in two intermediate coupled models with statistical atmospheres and simplified ecosystems (Timmermann and Jin 2002; R.-H. Zhang et al. 2009); and an increase in

amplitude is reported in two full GCMs (Lengaigne et al. 2007; Anderson et al. 2007), and in two intermediate coupled models with full ecosystems (Marzeion et al. 2005; Ballabrera-Poy et al. 2007). The implicit null hypothesis in the studies above is that light absorption through chlorophyll strengthens (or weakens) ENSO. It is difficult to compare these studies with each other, because the results from testing this hypothesis will depend on whether the control experiment has no chlorophyll at all, or spatially or temporally constant chlorophyll, and which mean concentration is assumed for the latter two.

It is clear, however, from all of these pioneering studies that chlorophyll matters for the mean and seasonal cycle of stratification and SST. Thus, since 2004 all Community Climate System Model (CCSM) versions use the observed climatology of chlorophyll to compute the shortwave absorption in the upper ocean (Danabasoglu et al. 2006, their appendix B). It is not clear from any of these studies, however, whether the changes in ENSO are due to changes in either the climatology or the ENSO–phytoplankton feedbacks. These changes are not only of scientific interest but also of societal importance, because if the latter is the case, then the current practice of climate modelers of using the observed chlorophyll climatology will misrepresent ENSO and could affect the forecasting of El Niño and La Niña events. It is then also likely that

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projections of global change scenarios are misleading, unless we understand how phytoplankton interacts with the physical system on climate-relevant scales. Therefore, the focus of the present study is on isolating the contribution of interannual chlorophyll variability to ENSO. It is shown that compared with a simulation using a fully interactive ecosystem, a simulation using identical prescribed chlorophyll climatology will have an increased tropical variability.

The next section provides a detailed description of the experiments, section 3 discusses the results, and a summary concludes this manuscript.

2. Description of model and experiment

The numerical experiment is performed using the National Center for Atmospheric Research (NCAR) CCSM version 3 (CCSM3), which consists of fully coupled atmosphere, ocean, land, and sea ice models; a detailed description can be found in Collins et al. (2006).

The ocean model has a zonal resolution that varies from 340 km at the equator to 40 km around Greenland and a meridional resolution that varies from 70 km at the equator to 40 km around Greenland and 350 km in the North Pacific. This spatially varying resolution is achieved by placing the north pole of the grid over Greenland, and it reflects the different relevant length scales of the two processes that we think are most important for maintaining a stable global climate: deep convection around Greenland and in the Arctic, and oceanic heat uptake at the equator. In the vertical there are 25 depth levels; the uppermost layer has a thickness of 8 m, the deepest layer has a thickness of 500 m. The atmospheric model uses T31 spectral truncation in the horizontal (approximately 3.75° resolution), with 26 vertical levels. The sea ice model shares the same horizontal grid as the ocean model, and the land model is on the same horizontal grid as the atmosphere. This setup (called T31 \times 3) has been developed specifically for long paleoclimate and biogeochemistry integrations, and its performance is described in detail by Yeager et al. (2006). The most significant difference between the present model setup and the one described in Yeager et al. (2006) is the new atmospheric convection scheme, which leads to significant improvements in the simulation of ENSO (Neale et al. 2008). The changes to the OGCM are a stratification-dependent thickness diffusivity (Danabasoglu and Marshall 2007), which improves the equatorial thermocline; a reduced viscosity (Jochum et al. 2008), which improves the equatorial surface currents; and an increased background diapycnal diffusivity, which stabilizes the meridional overturning circulation (Jochum 2009).

The CCSM biogeochemistry (BGC) model is the result of combining the upper-ocean ecosystem model of Moore et al. (2002) with the biogeochemistry module of Doney (2001). It includes the nutrients nitrate, ammonium, phosphate, iron, and silicate; four phytoplankton groups; one class of zooplankton; dissolved organic matter; and sinking particulate detritus; and it has been extensively tested within CCSM3 (Moore et al. 2004). Chlorophyll synthesis is regulated by the balance between photosynthetic carbon fixation and light absorption; it is also dependent on nitrogen uptake (for details see Geider et al. 1998).

The list of parameters and observable variables in the BGC model is extensive, and optimizing a global simulation requires considerable resources and skill (e.g., Gnanadesikan et al. 2004; Doney et al. 2009). Moreover, observations of BGC variables are often patchy, and available time series are often too short to address interannual variability. Tropical Pacific chlorophyll, the present focus, also suffers from observational uncertainties (Ballabrera-Poy et al. 2003; Carr et al. 2006) and large interannual-to-decadal variability (Murtugudde et al. 1999; Strutton et al. 2008; Rodgers et al. 2008), but it can be observed by satellite (McClain et al. 2004), and the record with the current Sea-viewing Wide Field-of-View Sensor (SeaWiFS) is now more than 10 yr long. To account for the challenges of optimizing the ecosystem model and the uncertainties in the observations, we discuss two sets of experiments: one set with tropical surface chlorophyll concentrations that are larger than those in the observations, and one with concentrations that are smaller. This approach should provide an estimate of the uncertainty associated with ecosystem modeling. A recent intercomparison showed most of the commonly used OGCM have similar physical subgrid parameterizations and biases (Griffies et al. 2009), which suggests that the present result—interactive phytoplankton reduces ENSO amplitude—is reproducible with other GCMs, but this still has to be tested in at least one other GCM.

Four fully coupled experiments are performed, and all of the results are based on the last 250 yr of the simulations. LOWFULL is a 500-yr integration with the fully interactive BGC model, LOWCLIM is a 290-yr integration (years 211–500) without the BCE model but with the chlorophyll climatology of LOWFULL computed from the years 191–210. HIGHFULL (HIGHCLIM) is identical to LOWFULL (LOWCLIM), except for a different sediment source of iron, which leads to higher chlorophyll concentrations in the tropical Pacific (Fig. 1). It should be noted that satellite observations of chlorophyll are based on light reflected mostly from the surface but also of a small fraction of light backscattered from deeper parts of the mixed layer. In the present simulations the

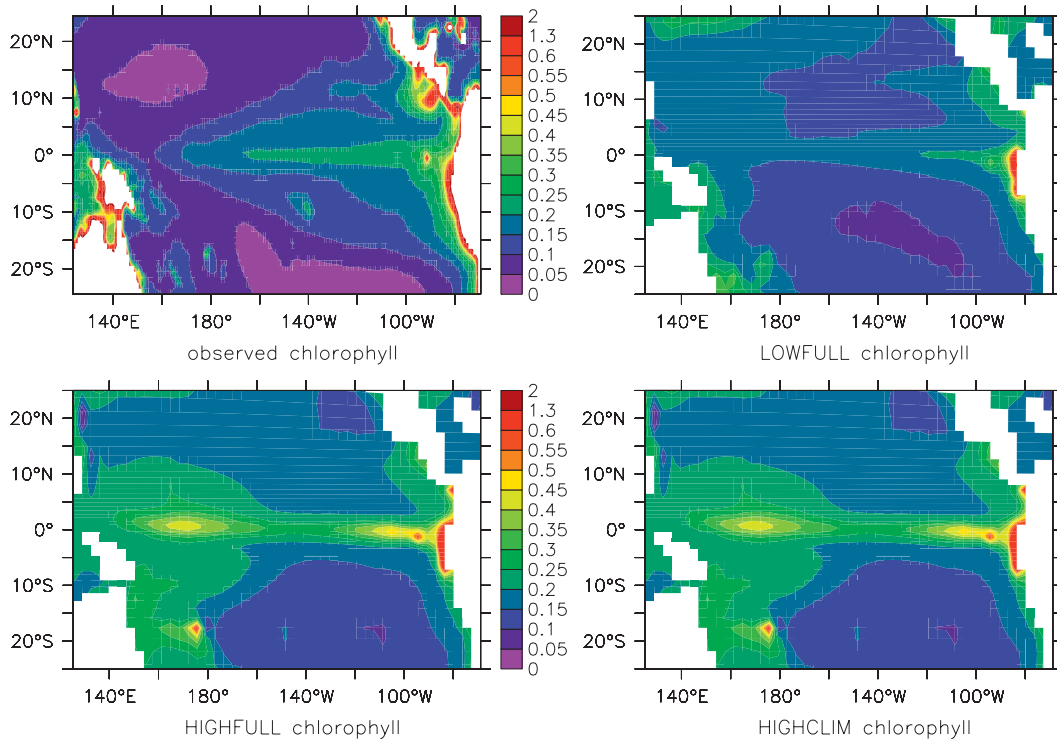


FIG. 1. Mean tropical Pacific surface chlorophyll concentrations (mg m^{-3}) based on years 1998–2007 in (top left) SeaWiFS (McClain et al. 2004), (top right) LOWFULL, (bottom left) HIGHFULL, and (bottom right) the prescribed concentration in HIGHCLIM.

observations are compared to the average chlorophyll concentration over the upper 30 m. Other depth ranges, like the surface layer, the mixed layer, or the depth by which shortwave radiation is reduced to a particular fraction of the surface value, can be argued for, but in the present set of experiments the respective values are almost identical in the tropical Pacific.

The average annual mean observed chlorophyll concentration in the Niño-3 region (the area from 5°N to 5°S and from 150° to 90°W) is 0.20 mg m^{-3} , while in LOWFULL it is 0.18 mg m^{-3} , and in HIGHFULL it is 0.25 mg m^{-3} , with standard deviations (std dev) of 0.02, 0.01, and 0.04 mg m^{-3} , respectively. This information suggests that for the direct equatorial response, the two sets of experiments will provide lower and upper bounds for the ENSO response. Another way to assess the fidelity of the BGC model is to compare variability of new production along the equator. For 1992–2000 Turk et al. (2001b) estimate the integrated new production between 1°S and 1°N , and between 140°E and 100°W to be $1.6 \times 10^{12} \text{ mol yr}^{-1}$ nitrogen, with a minimum of $1.2 \times 10^{12} \text{ mol yr}^{-1}$ during the 1997/98 El Niño and a maximum of $2.1 \times 10^{12} \text{ mol yr}^{-1}$ during the following La Niña. The mean for LOWFULL is $1.3 \times 10^{12} \text{ mol yr}^{-1}$, with extrema of 1.1×10^{12} and $1.6 \times 10^{12} \text{ mol yr}^{-1}$, and

for HIGHFULL the mean is $2.1 \times 10^{12} \text{ mol yr}^{-1}$, with extrema of 1.6×10^{12} and $3.2 \times 10^{12} \text{ mol yr}^{-1}$. Thus, like the chlorophyll concentration, the new production suggests that LOWFULL and HIGHFULL provide a lower bound and an upper bound for the ENSO response. For the interpretation of the experiments it is important that the tropical Pacific chlorophyll concentrations do not exhibit large drifts during the final 250 yr of the experiments. The drift can be estimated by comparing the mean chlorophyll concentrations in HIGHFULL for the last 250 yr with those of HIGHCLIM (bottom of Fig. 1, based on yr 191–210 of HIGHFULL); both are mostly within a few percent of each other.

The choice of using different iron sources to tune the model may appear unusual and warrants a short justification. Plankton growth in the tropical Pacific is iron limited (e.g., Martin et al. 1994), and using two different sediment iron sources is a way to control tropical Pacific chlorophyll concentrations without affecting the physical system or the BGC in other parts of the world. The key difference between the iron fields of LOWFULL and HIGHFULL is the iron source off New Guinea (Fig. 2). High dissolved iron concentrations are observed in this western Pacific region (Mackey et al. 2002), and the additional iron in HIGHFULL is swept east in the

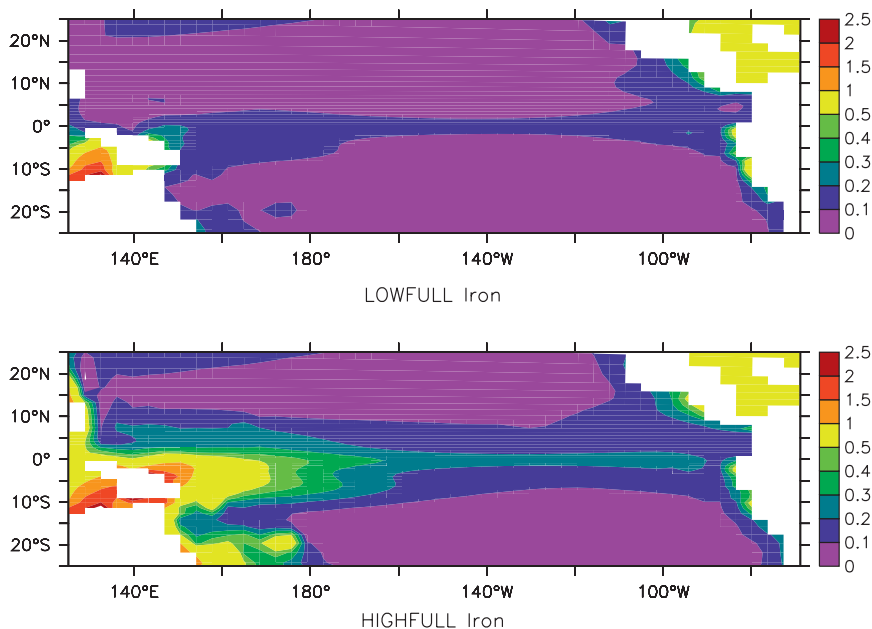


FIG. 2. Simulated concentrations ($\mu\text{mol m}^{-3}$) of dissolved inorganic iron in the upper 150 m of (top) LOWFULL and (bottom) HIGHFULL.

Equatorial Undercurrent and fertilizes the equatorial Pacific. The two different iron source fields have not been developed with the present experiment in mind. Instead, they emerged as part of research into locations and impacts of sediment iron sources (Moore and Braucher 2008).

3. Results

In all of the experiments the shape of the ENSO spectrum and the relative seasonal distribution of its variance are similar and not statistically different from each other; thus, only the spectra of HIGHFULL and HIGHCLIM are shown for illustrative purposes (Fig. 3). Significant differences can be found, however, in the ENSO amplitude. The std dev of Niño-3 SST for LOWFULL is 0.93°C and for LOWCLIM it is 1.02°C , for HIGHFULL it is 0.90°C and for HIGHCLIM it is 0.99°C . The statistical significance of the differences is established with a variant of the nonparametric Mann-Whitney test for variances (von Storch and Zwiers 1999). The 250-yr record of each experiment is divided into 10–14 subrecords, and it is found that the std dev of Niño-3 SST in

- LOWCLIM is larger than in LOWFULL at 90% significance (A),
- HIGHCLIM is larger than in HIGHFULL at 98.5% significance (B),
- LOWFULL is larger than in HIGHFULL at 90% significance (C), and

- LOWCLIM is larger than in HIGHCLIM at 95% significance (D).

The significance levels are independent of the number of subrecords (as long as they are between 10 and 14; other values have not been tested).

These results lead to the following three questions: First, what explains the amplitude reduction between the two HIGH cases and their respective LOW counterparts (C and D)? Second, what explains the amplitude reduction between the two FULL cases and their respective CLIM cases (A and B)? And, last, why is the amplitude difference between HIGHFULL and HIGHCLIM the same as between LOWFULL and LOWCLIM, despite the larger equatorial chlorophyll concentrations in the HIGH cases?

The first question is ultimately asking for a general theory of ENSO amplitude, which to our knowledge is still not available. Furthermore, in the present version of CCSM, ENSO is sensitive to even small changes in SST (Jochum et al. 2009). Still, some insight can be gained by analyzing the mean fields. Both CLIM cases are run with prescribed but different chlorophyll (and therefore different near-surface stratification), so they can be looked at as sensitivity studies of ENSO with different prescribed vertical mixing (Gnanadesikan and Anderson 2009). Indeed, the equatorial stratification in and just below the mixed layer is stronger in HIGHCLIM than in LOWCLIM, which coincides with slightly reduced cross-equatorial SST gradients, shifted precipitation fields

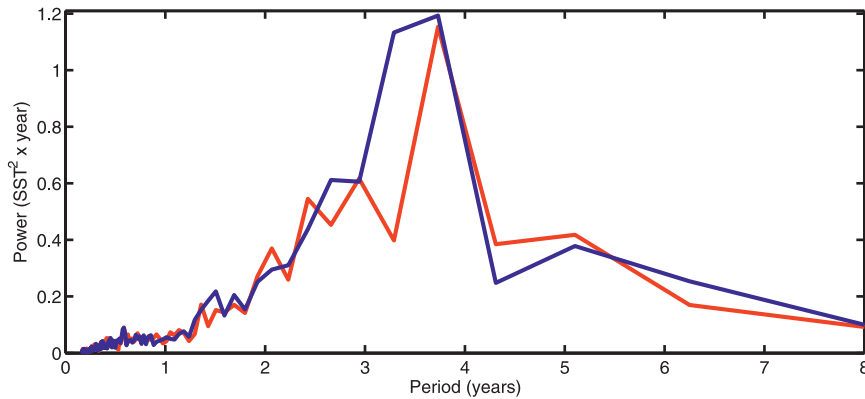


FIG. 3. Power spectrum for Niño-3 SST anomalies ($^{\circ}\text{C}$) for HIGHFULL (red) and HIGHCLIM (blue). The area under the lines integrated across all frequencies yields the total variance.

(Fig. 4), and weaker upwelling favorable cross-equatorial winds (not shown). The model is fully coupled, so it is difficult to establish causality unequivocally; however, because the only difference between the HIGHCLIM and LOWCLIM setups is the chlorophyll climatology, the change to the mean climate has to come through the connection between chlorophyll, SST, and surface wind stress. Here, this is not a local direct connection as in the idealized study by Timmermann and Jin (2002) that requires a warming of the eastern equatorial SST; instead, the response has to be remotely forced (as in, e.g., Schneider and Zhu 1998) because the SST is cooler despite increased surface chlorophyll and reduced upwelling. The equatorial cooling can be traced back to off-equatorial subsurface cooling that is induced by the increased primary production in the western Pacific between 20° and 10°S (not shown, but see Fig. 1 for the differences in chlorophyll). This point suggests that the increased stratification is not caused directly by increased equatorial chlorophyll concentrations but is due to reduced upwelling-favorable winds, which are ultimately caused by increased off-equatorial chlorophyll concentrations (as in Anderson et al. 2007).

The second question is best answered in reference to the coupled model studies mentioned in the introduction. The connection between these previous studies and the present experiments is not straightforward, because each study uses a different GCM and different control experiments. A robust result, however, is that the direct heat-trapping effect of chlorophyll leads to a reduction of ENSO amplitude (as in Timmermann and Jin 2002; Wetzels et al. 2006; R.-H. Zhang et al. 2009), unless the phytoplankton weakens the seasonal cycle in the eastern tropical Pacific, in which case ENSO becomes stronger (as in Marzeion et al. 2005; Lengaigne et al. 2007; Ballabrera-Poy et al. 2007). The inverse connection between the strengths of ENSO and the seasonal cycle of the east

equatorial Pacific SST is a robust feature of GCMs (e.g., Chang et al. 1995; Guilyardi 2006); for the present analysis, however, this connection is of little consequence because the comparisons conducted here are between the FULL and CLIM cases, which by construction have the same seasonal cycle of chlorophyll and a very similar seasonal cycle for everything else (not shown). Thus, the present results are consistent with the first three studies above, in which each predicts a weakening of ENSO of approximately 10% for fully prognostic chlorophyll. The heat-trapping effect on ENSO has been observed and described by Strutton and Chavez (2004): during El Niño, equatorial upwelling and therefore phytoplankton nutrient supply is reduced and the mixed layer depth is increased, which leads to reduced chlorophyll concentrations and biologically induced heating; the reverse takes place during La Niña.

While the present results can be explained by this mechanism, the question remains as to why the amplitude of the SST response is independent of the mean chlorophyll concentration or its std dev in the Niño-3 region. In both sets of LOW and HIGH experiments, the full ecosystem reduces the std dev of Niño-3 SST by 9%. This point suggests that in HIGHFULL the ecosystem not only acts to damp the local Niño-3 SST anomalies but that it also amplifies ENSO dynamically so that the two effects—increased thermodynamic damping and dynamical amplification—cancel each other.

The understanding of this dynamical amplification will depend on the guiding ENSO theory, for which there are many choices (Wang and Picault 2004). Since Neale et al. (2008), though, ENSO in CCSM3 can be described as a combination of a “delayed oscillator” and “a series of events.” In the latter, individual El Niño events are triggered by strong westerly wind anomalies in the western Pacific (e.g., Keen 1982), and every event is independent of the previous one; in the former, ENSO is a real

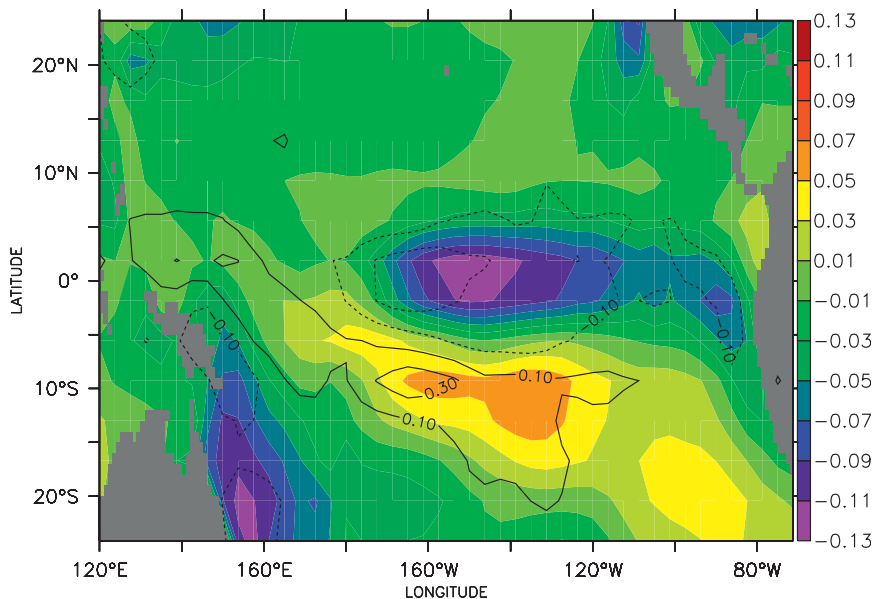


FIG. 4. Difference between mean SST ($^{\circ}\text{C}$; color) and precipitation (mm day^{-1} ; contours) between HIGHCLIM and LOWCLIM.

oscillation, with the memory being in the equatorial Kelvin and the off-equatorial Rossby waves (e.g., Zebiak and Cane 1987). Compared to the high-resolution version of CCSM that is discussed in Neale et al. (2008), which simulates realistic westerly wind anomalies, the present coarse-resolution version has relatively weak westerly wind anomalies and is therefore more dominated by “delayed oscillator dynamics” (Jochum et al. 2009). A necessary component of this regime is the generation of off-equatorial Rossby waves by ENSO related off-equatorial wind anomalies (Neale et al. 2008).

Thus, ENSO strength cannot only be measured through SST but also through thermocline depth variability (see also Meinen and McPhaden 2000). Figure 5 (top) shows a proxy for the latter: the interannual variability of the 20°C isotherm depth. The variability is large not only in the eastern equatorial Pacific along the Kelvin wave path but also in the off-equatorial western Pacific centered near the date line at 7°S and 6°N , where the ENSO-induced Rossby waves are active. It is noteworthy that there is relatively little Rossby wave activity in the eastern part of the basin, because the waves do not draw their energy from reflected equatorial and coastal Kelvin waves but from the anomalous surface wind stresses in the central Pacific. This process is discussed and illustrated in detail by Neale et al. (2008) but can also be seen in Fig. 5 (bottom): El Niño conditions lead to westerlies on the equator and easterlies near 15°N/S (as indicated by positive and negative correlations, respectively), which leads to upwelling Rossby waves in between. Comparing the top panels of Fig. 5 then shows that, in the western

basin, thermocline depth variability is 5%–10% stronger in HIGHFULL than in LOWFULL. This stronger subsurface variability, which is indicative of stronger planetary wave activity, is caused by a stronger response of the off-equatorial wind field between 20° and 10°S and between 10° and 20°N to equatorial SST anomalies [Fig. 5 (bottom)].

The detailed process by which this amplification happens is more difficult to come by. The location of the most important chlorophyll response is elusive because of the nonlocal connection between tropical SST and wind stress (e.g., Gill 1980; Barsugli and Sardeshmukh 2002). Thus, in the framework of the present coupled model studies it is not possible to unequivocally determine this location and hence the process. The results do reveal, though, that the relatively stronger off-equatorial anticorrelation of the winds that is seen in HIGHFULL (compared to LOWFULL) is not seen in the two CLIM experiments, and neither is the east–west difference in change of thermocline variability (not shown). This point suggests that the equatorial phytoplankton population has to be large *and* interactive to lead to dynamical amplification. However, especially in the warm pool, the HIGH simulations have a less realistic chlorophyll distribution than the LOW simulations, which is not only evident from the comparison of the surface concentrations (Fig. 1) but also from an analysis of the subsurface concentrations in the equatorial western Pacific (between 145° and 165°E). There, the mean subsurface chlorophyll maximum in the LOW simulations is at a depth of 60 m with a magnitude of 0.3 mg m^{-3} , one-third of the surface

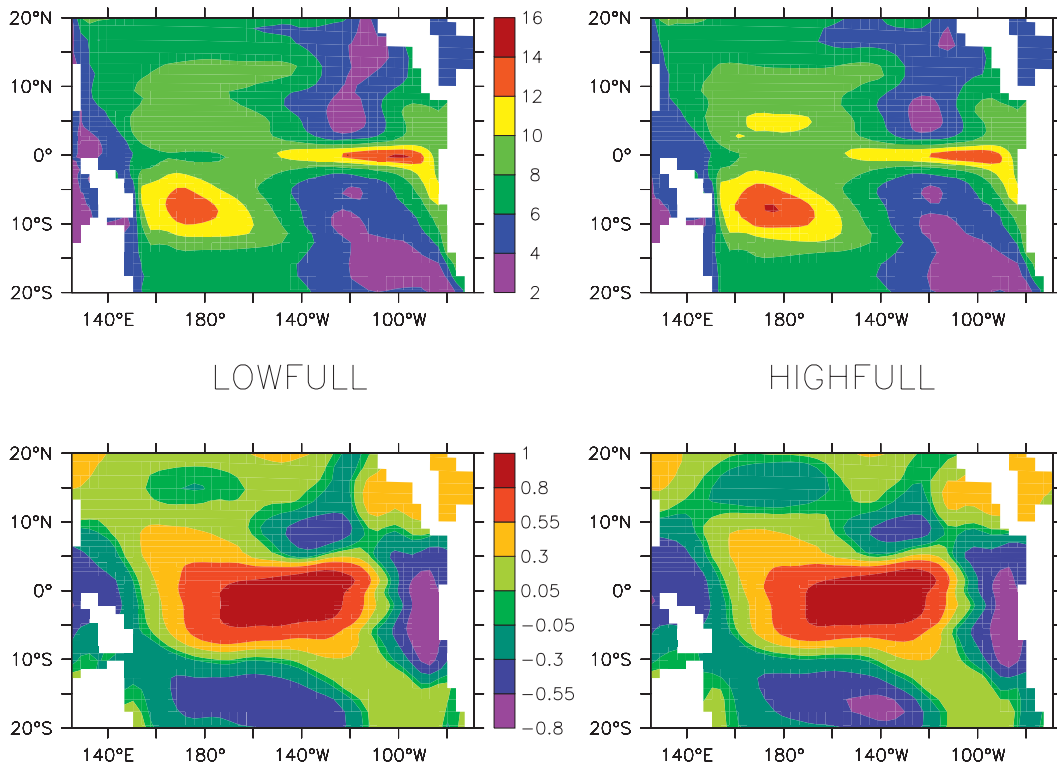


FIG. 5. (top) Std dev of the 20°C isotherm depth (contour interval: 2 m): (left) LOWFULL and (right) HIGHFULL. (bottom) Correlation between SST anomalies at 0°, 110°W and zonal wind stress anomalies in the tropical Pacific: (left) LOWFULL and (right) HIGHFULL. All correlations are based on 250 yr of annual means with zero lag.

concentration. This structure is consistent with the in situ observations by Turk et al. (2001a). The HIGH simulations, on the other hand, have no distinct chlorophyll maximum in the equatorial western Pacific but have the same chlorophyll concentration throughout the upper 60 m (not shown). Given the lesser realism of the HIGH simulations, their results should only be viewed as what they were designed to be: an upper bound for the impact of interactive phytoplankton. The dynamical feedback that is uncovered through the comparison between LOWFULL and HIGHFULL is then mainly an explanation of the model results, and not the postulate of a real world process.

After frequency, which is similar in all experiments, and amplitude, which is the focus of this section, skewness of the Niño-3 SST is occasionally used as a third metric to characterize ENSO (Burgers and Stephenson 1999). It quantifies to what extent warm anomalies (El Niños) are larger than cold anomalies (La Niñas), and Rodgers et al. (2004) link it to decadal variability. It is relevant in the present context, because Timmermann and Jin (2002) speculate that interactive phytoplankton could affect skewness, because the chlorophyll concentration has natural limits: it cannot sink below zero during El Niño, and presumably during La Niña there is also a

limit beyond which the chlorophyll concentration stops growing. Whether the tropical Pacific operates in such extreme regimes, however, is not known. The only quantitative model study that the authors are aware of is done with forced OGCM simulations, and there the difference between no chlorophyll at all, and fully interactive chlorophyll is, among other things, an increase in Niño-3 SST skewness from -0.63 to 0.14 (Löptien et al. 2009). For orientation, the observed skewness is 0.86 (T. Zhang et al. 2009). The source of the increase in the OGCM study is difficult to determine, because adding chlorophyll in the OGCM also leads to a cooling of the mean Niño-3 SST by approximately 1°C . In the present experiments the skewness in the FULL simulations is not significantly different from the skewness in the respective CLIM counterparts. Thus, in the present GCM, flow-dependent phytoplankton does not affect skewness. Changing the mean chlorophyll concentration, however, does affect it, and for the sake of completeness the numbers are reported here. Increasing the background iron, and therefore chlorophyll concentrations, leads to a slight increase in skewness from 0.25 (± 0.08) in LOWFULL to 0.38 (± 0.06) in HIGHFULL (the numbers in brackets are the respective std dev based on five 50-yr subintervals). Thus, the connection between the slightly cooler Niño-3 mean

SST in the HIGH simulations (Fig. 4) with a slightly larger skewness is consistent with the forced OGCM findings of Löptien et al. (2009) that a much cooler mean Niño-3 SST leads to a much larger skewness. It is beyond the scope of the present study to discuss skewness in more detail, but a detailed account of processes that affect skewness in CCSM is provided by T. Zhang et al. (2009).

4. Summary and discussion

Four multicentury integrations of a current coarse-resolution version of CCSM are used to investigate the effect of phytoplankton on ENSO. It is found that a simulation with a fully active ecosystem has a 9% smaller ENSO amplitude than one with the prescribed but identical climatological chlorophyll values. This holds true for mean surface chlorophyll concentrations that are larger, as well as smaller, than those observed. Therefore, it is not sufficient for studies of tropical variability to represent the effect of phytoplankton by simply including the observed annual cycle of surface chlorophyll. Because the full ecosystem is a rather expensive component of the full GCM (especially in high-resolution GCMs), it is desirable to enhance the included observed climatology with either a statistical model that on interannual time scales connects SST with chlorophyll, or a reduced ecosystem model that only provides chlorophyll as prognostic variable. Future research will have to determine which of these two options is the most promising, but it appears that for applications in different climates (either ice age or global warming scenarios) the simplified prognostic model is the better approach, because the present-day empirical relationship may break down.

The present results appear to be in contradiction to Löptien et al. (2009) who find that interactive phytoplankton *increases* ENSO amplitude by 5%. While two particular differences in the experimental design could account for this difference (uncoupled versus coupled, annual mean chlorophyll as the control versus seasonally varying chlorophyll as the control), the statistics of the present simulations suggests that there might not be a contradiction at all. Their 55-yr integration can be thought of as being one particular interval of ENSO activity in which interactive phytoplankton may indeed have had an amplifying effect. There is, however, substantial natural variability in ENSO activity on decadal-to-centennial time scales [Wittenberg (2009), see also Stevenson et al. (2010, manuscript submitted to *J. Climate*) for the present model, and Garcia-Herrera et al. (2008) for observations], and in each of the two sets of present experiments there are only two 50-yr intervals in which the FULL variability is larger than in the lowest variability interval of the CLIM simulations. Thus, while

the experimental design in Löptien et al. (2009) could explain the contradiction, it is possible that 50- or even 100-yr intervals are not sufficiently long to assess the impact of interactive phytoplankton.

The structure of satellite-based chlorophyll data in any given week in the tropical Pacific is often dominated by variability on scales of several hundred kilometers resulting from tropical instability waves (Yoder et al. 1994). The zonal resolution in the present model configuration does not allow detection these features, which raises the question of how representative the present results are for the real world, or even for the coming next generation of GCMs, with eddy-resolving OGCMs that include ocean ecosystem models. The existence of tropical instability waves has been known for more than 30 yr (e.g., Düing et al. 1975), but the complexity of their structure (e.g. Kennan and Flament 2000) still leads to surprising discoveries (Jochum et al. 2007). It is not surprising then that their net impact on equatorial chlorophyll is still debated: earlier observations suggest that they increase mean chlorophyll (Barber et al. 1996), whereas the most recent observations suggest a slight reduction of mean chlorophyll during periods of strong instability wave activity (Evans et al. 2009). The only modeling study that the present authors are aware of suggests that the presence of instability waves reduces mean the concentration of equatorial chlorophyll by approximately 5% (Gorgues et al. 2005). Given that the 40% difference in the equatorial chlorophyll concentrations in the current set of experiments did not lead to a difference in the ENSO–phytoplankton feedback, one may speculate that a future generation of eddy-resolving ocean models will not show a vastly different sensitivity to interactive phytoplankton either.

In the opinion of the authors, the present demonstration of a two-way feedback between ENSO and phytoplankton should lead to a closer collaboration between physical oceanographers and marine biologists. It is current practice, not only at NCAR, that models of ocean physics are developed without much thought about marine ecosystems. Similarly, models of marine ecosystems are developed using the output of OGCMs without the consideration of possible feedbacks. This situation is reminiscent of the state of AGCM and OGCM developments before the 1982/83 El Niño, and much time has been spent since then bringing the respective communities closer together. One may argue that the present 9% change in amplitude is smaller than the current amplitude bias in many GCMs (Guilyardi 2006), but the biases will be reduced in time. Furthermore, the nonlinear behavior that has been observed in some El Niño events suggests that for predictability purposes even seemingly minor processes can become important (Vecchi et al. 2006). Apart

from ENSO predictability, the present results are also relevant to the global warming discussion. One of the regional effects of global warming that has relatively solid theoretical, modeling, and observational support is the warming of the cold tongue and a slow down of the Walker circulation (e.g., Vecchi and Soden 2007). Considering the robust connection between SST and chlorophyll in the cold tongue, one would expect that phytoplankton ameliorates this trend. This idea, however, still has to be tested.

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