

A role for ocean biota in tropical intraseasonal atmospheric variability

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[1] We propose that temporal variations within the marine plankton system can induce intraseasonal variations in sea surface temperature (SST) through the effect on solar penetration due to chlorophyll and other optically active organic components. Sensitivity studies with a simple model suggest that these small oscillations in SST may stimulate radiative-convective oscillations in the atmosphere which amplify them and thus induce or modulate significant variability in the coupled system. Long term bio-optical measurements in the Western Pacific, where satellite time series are degraded by clouds, would provide a test of our theory and would improve our understanding of the heat balance in this climatically important region. *INDEX TERMS*: 0315 Atmospheric Composition and Structure: Biosphere/atmosphere interactions; 4815 Oceanography: Biological and Chemical: Ecosystems, structure and dynamics; 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 3314 Meteorology and Atmospheric Dynamics: Convective processes; 4855 Oceanography: Biological and Chemical: Plankton. **Citation**: Gildor, H., A. H. Sobel, M. A. Cane, and R. N. Sambrotto, A role for ocean biota in tropical intraseasonal atmospheric variability, *Geophys. Res. Lett.*, 30(9), 1460, doi:10.1029/2002GL016759, 2003.

1. Introduction and Model Description

[2] The tropical intraseasonal oscillation (ISO) is a prominent mode of atmospheric variability in the tropics. It is quasi-periodic with a typical period of 40–50 days [Madden and Julian, 1994]. The ISO interacts with the El Niño-Southern Oscillation [Zhang *et al.*, 2001], the monsoons, and extratropical weather [Ferranti *et al.*, 1990], and so is of global importance. Despite many attempts to explain the origin and characteristic of the ISO [see Waliser *et al.*, 1999, and references therein] there is no widely accepted explanation for these waves and atmospheric models fail to simulate them realistically.

[3] Based on sensitivity studies with a simple model, we suggest that oscillations within the marine plankton system, which can affect the penetration of solar radiation through the ocean upper layer and hence influence sea surface temperature, may be powerful enough to initiate or modulate intraseasonal atmospheric variability in the tropics. This argument is broadly in accord with recent theoretical and modeling studies [Wang and Xie, 1998; Watterson, 2002] suggesting a

significant role for air-sea interaction in ISO dynamics, but adds a new, independent source of variability, the biota.

[4] The simple model consists of an ocean mixed layer, including ocean biota, and a simple atmospheric model, all at a single horizontal space point. The marine plankton model, based on Edwards and Brindley [1999], is a fairly standard implementation of a three compartment [phytoplankton (P), zooplankton (Z), nutrient (N)] surface ocean ecological system within a homogeneous mixed layer ocean with a constant depth of 25 m. All pools are expressed in N equivalent. The governing ordinary differential equations are:

$$\frac{dN}{dt} = -\nu \frac{N}{e+N} P + mP + \frac{\beta \lambda P^2}{\mu^2 + P^2} Z + \gamma qZ + k(N_0 - N)$$

$$\frac{dP}{dt} = \nu \frac{N}{e+N} P - mP - \frac{\lambda P^2}{\mu^2 + P^2} Z$$

$$\frac{dZ}{dt} = \frac{\alpha \lambda P^2}{\mu^2 + P^2} Z - qZ$$

[5] Phytoplankton uptake of nutrient is represented by the Michaelis-Menten function $\frac{N}{e+N}$ where e is the half saturation constant and ν the maximum growth rate. We assume no phytoplankton growth limitations due to light [McClain *et al.*, 1999] and no self-shading by phytoplankton as the amount of chlorophyll is always relatively low. We also neglect the sinking of phytoplankton, reflecting their small size in this region [McClain *et al.*, 1999] and mixing of phytoplankton with the layer below. Mortality and respiration of phytoplankton are included in mP . Consumption of phytoplankton by zooplankton is parameterized by a sigmoidal Holligan type III term, $\frac{\lambda P^2}{\mu^2 + P^2} Z$, with maximum zooplankton grazing rate λ and Z grazing half-saturation constant μ . Although the half-saturation value for grazing used here is less than that used in Edwards and Brindley [1999], it is justified by the fact that most of the models summarized in Edwards and Brindley [1999] were based on macro-zooplankton grazing, while in the tropical Pacific the micro-zooplankton are mainly responsible for keeping phytoplankton growth in check. Part α of the grazing represents growth efficiency of the zooplankton, and part β represents zooplankton excretion which is regenerated spontaneously back into the nutrient pool while the rest is lost from the system as faecal pellets. Zooplankton mortality is linear with q as the mortality rate, and a fraction γ of it is regenerated back to the nutrient pool. The nutrient concentration below the upper layer is N_0 , and k is the mixing rate.

[6] For a wide regime of parameters, including those specified in Table 1, the model exhibits self-sustained oscillations with a period of approximately 60 days. The oscillations in the amount of phytoplankton and zooplankton are intrinsic to the ecological system (panels (c) and (d) of Figure 1), and in our model are the result of predator-prey interaction [Murray, 1989]. Predator-prey oscillations of

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Table 1. Model Parameters

Parameter	Symbol	Value
Ecological Model		
Maximum growth rate	ν	2.5, day ⁻¹
Half-saturation constant for N uptake	e	1.1 mmol N m ⁻³
Cross mixed-layer base exchange rate	k	0.03 day ⁻¹
Higher predation on Z	q	0.11 day ⁻¹
P mortality and respiration rates	m	0.13 day ⁻¹
N concentration below the mixed layer	N_0	0.6 mmol N m ⁻³
Z growth efficiency	α	0.25
Z excretion fraction	β	0.33
Regeneration of Z predation excretion	γ	0.7
Maximum Z grazing rate	λ	0.6 day ⁻¹
Z grazing half-saturation coefficient	μ	0.004 g N m ⁻³
Atmospheric model		
Radiative equilibrium temperature	T_e	-60 K
Atmospheric temperature	T	-10 K
Radiative relaxation time	τ_R	30 days
Convective relaxation time	τ_c	0.3 days
Evaporation time scale	τ_E	12 days
Cloud-radiative feedback	r	0.2-0.3
Vertical mean of the moisture basis function	\hat{b}	0.455
Ocean mixed layer model		
Short wave radiation at the top of the atmosphere	S_0	240 W m ⁻²
Long wave radiation	L	40 W m ⁻²
Heat capacity of the mixed layer	C	1.053e8 J kg ⁻¹ K ⁻¹

similar periodicity have been predicted by other ecological models ranging from two to seven compartments [Steele and Henderson, 1992; Fasham, 1993], and have been observed in the North Atlantic [Williams, 1988], in many lakes [McCauley and Murdoch, 1987] and microcosm experiments [McCauley et al., 1999]. In all of these, the period of the oscillation is of the order of a few tens of days. Similarly, for a wide range of model parameters, the period of the predator-prey oscillations is few tens of days [Edwards and Brindley, 1999]. Long term measurements of phytoplankton levels in the Western Pacific are not available for comparison.

[7] The atmospheric model is a zero-dimensional reduction of the Neelin-Zeng Quasi-Equilibrium Tropical Circulation Model (QTCM) [Neelin and Zeng, 2000], with the atmospheric temperature fixed, as is arguably appropriate for tropical single-column simulations [Sobel and Bretherton, 2000]. We also neglect horizontal advection of moisture, taking the horizontal moisture convergence to be equal to the mass convergence times the local specific humidity. The equations are:

$$M_S \delta = Q - R \quad (1)$$

$$\hat{b} \frac{\partial q}{\partial t} - M_q \delta = E - Q \quad (2)$$

$$Q = \frac{q - T}{\tau_c}, \quad q > T; \quad Q = 0, \quad q \leq T \quad (3)$$

$$R = \frac{T - T_e}{\tau_R} - rQ \quad (4)$$

[8] (1) originates as an equation for the atmospheric temperature T , but the rate of change $C_p dT/dt$ has been dropped. Q is convective heating, or equivalently precipitation, R radiative cooling (assumed positive), E surface evaporation, δ horizontal divergence in the upper tropo-

sphere and q the latent heat of vaporization of water times the specific humidity. Q , E and R have the units J kg⁻¹s⁻¹. The velocity field, atmospheric temperature and q implicitly multiply fixed structure functions in the vertical coordinate. We use the structure functions, reference profiles, and constants and coefficients derived from them, from the version 2.2 of the QTCM; see Table 1 of Neelin and Zeng [2000]. We have neglected surface sensible heat flux since it is small compared to evaporation over tropical oceans. M_S and M_q are the dry static stability and gross moisture stratification, taken to be linearly increasing functions of T and q ; here M_S is fixed since T is. M_q varies with q but these variations are not a primary factor in the dynamics here. The constant \hat{b} derives from the projection of the advection term on the structure functions of the horizontal velocity. The precipitation is parameterized using the simplified Betts-Miller scheme [equation (3)] with τ_c the convective time scale. τ_R is a radiative time scale, T_e a radiative equilibrium temperature, and r is a nonnegative dimensionless cloud-radiative feedback parameter. Our best guess for r is 0.2, with possible uncertainties up to 50% [Bretherton and Sobel, 2002]. We assume the troposphere is always cooled rather than heated by radiation in the vertical mean, so if R based on the above expression becomes negative, it is set to zero. The evaporation is parameterized by the bulk formula $E = (q_s^*(T_s) - q_s)/\tau_E$ where q_s^* is the saturation specific humidity at the sea surface temperature T_s , and q_s is related linearly to q by the vertical structure function and reference profile and τ_E is the evaporation time scale. Taking this parameter fixed is equivalent to fixing the surface wind speed, which is an important limitation of the model, although an acceptable one for our purpose.

[9] The temperature of the ocean mixed layer is governed by:

$$C \frac{\partial T_S}{\partial t} = S - L - E \quad (5)$$

$$S = 0.53S^* + 0.47S^* \left(1 - e^{-(0.027 + 0.0518^* Ch^{0.428})^* D} \right) \quad (6)$$

where C is the heat capacity of the mixed layer, S represents the short-wave radiation absorbed within the mixed layer, and L (which is fixed) represents the combined effects of

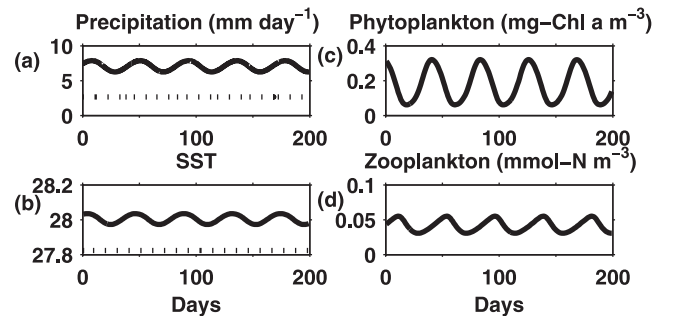


Figure 1. Stimulation of intraseasonal atmospheric oscillations by the ocean biota, for $r = 0.2$. Evolution of precipitation (a) and SST (b) when ocean biota do not affect light penetration (dotted lines) and when ocean biota do affect light penetration (solid lines). The two right hand panels show the prey-predator oscillations within the phytoplankton-zooplankton-nutrient system. (c) phytoplankton and (d) zooplankton.

longwave radiation and ocean heat transport. S will be modified by the cloud-radiative feedback and by the amount of chlorophyll in the mixed layer. Assuming that the short wave radiation at the top of the atmosphere is S_0 , the amount of radiation at the ocean surface, taking into account the cloud-radiative feedback, is $S^* = S_0 - rQ$. By using here the same r as in (4), the reflection of shortwave radiation and greenhouse trapping of infrared radiation by clouds are taken equal, as is approximately observed [Ramanathan *et al.*, 1989]. Of the solar radiation reaching the ocean surface, approximately half (53%) is absorbed within the upper few meters, while the rest is attenuated at different rates, dictated mainly by the amount of chlorophyll [Morel, 1988]. We calculate S in (6) based on the amount of radiation at the ocean surface and the amount which penetrates the base of the mixed-layer [Morel, 1988] where D is the mixed layer depth and Chl the amount of chlorophyll (in mg m^{-3}). In the Western Pacific Warm Pool, where the mixed layer is shallow [Lukas and Lindstrom, 1991; Lewis *et al.*, 1990] and light can penetrate below the mixed layer, the effect on ocean sea surface temperature can become significant [Siegel *et al.*, 1995; Lewis *et al.*, 1990; McClain *et al.*, 1999].

2. Results and Discussion

[10] The coupled model exhibits oscillatory behavior over a wide range of parameters, and a detailed analysis of its behavior (without biology) will be published elsewhere (Sobel and Gildor, submitted, 2003). Through sensitivity studies, we illustrate that the model is, within a fairly broad and plausible range of the key parameters, sensitive to variations in SST of a few tenths of a degree. Calculations based on observations of biological processes in the Western Pacific Warm Pool suggest that SST variations of this magnitude are possible [Siegel *et al.*, 1995; Lewis *et al.*, 1990]. These variations can shift the model from a stable regime to an oscillatory one, or can modify the frequency within an oscillatory regime.

[11] The dotted lines in Figure 1 show atmospheric precipitation and SST as a function of time when the amount of chlorophyll does not affect light absorption within the upper mixed layer, i.e. assuming $Chl = 0$ in equation (6). The model reaches a steady state (dotted lines). Inclusion of the amount of chlorophyll calculated by the ecological model in equation (6) results in small amplitude oscillations in precipitation and SST (Figure 1, solid lines). With the cloud-radiative feedback parameter r increased from 0.20 to 0.25, the atmospheric model oscillates vigorously, with a time scale of 60 days and precipitation amplitude of about 25 mm day^{-1} (Figure 2, left panels, solid lines). Without the biology, the model converges to a steady state for this value of r (Figure 2, left panels, dotted lines). If r is further increased to 0.3, the model oscillates even without taking into account the effect of ocean biota (Figure 2, right panels, dotted lines), but the period of the oscillation is approximately 120 days. For this value of r , the biology modulates the time scale, reducing it to approximately 64 days (Figure 2, right panels, solid lines).

[12] To understand the model dynamics, consider first a case without biology. The surface evaporation E adjusts q towards the saturation value set by the SST. If this value is greater than T , at some point convection (measured by Q) will commence. Consistent with observations, the model exports moist static energy horizontally from convective

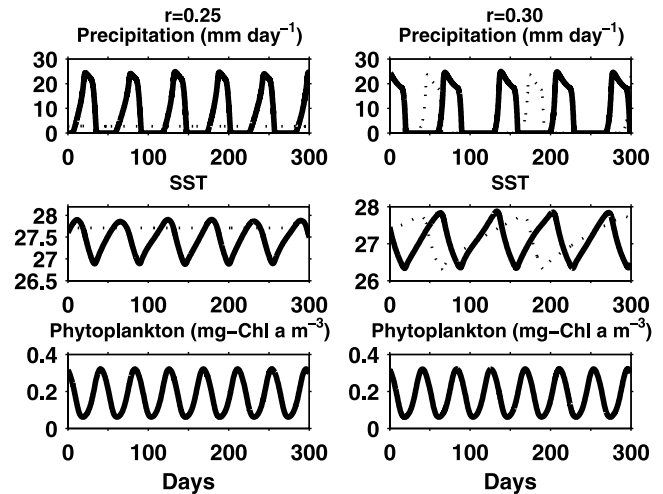


Figure 2. Stimulation of intraseasonal atmospheric waves by the ocean biota, for $r = 0.25$. Evolution of precipitation (a) and SST (b) when ocean biota do not affect light penetration (dotted lines) and when ocean biota do affect light penetration (solid lines). The lower panel (c) shows the oscillations in phytoplankton.

regions, requiring a compensating source (since the gross moist stability $M = M_s - M_q$ is positive [Neelin and Zeng, 2000]). The net source of moist static energy is $E-R$, so by reducing R the infrared cloud-radiative feedback amplifies the convection [Raymond, 2000], leading to an instability. The shortwave cloud feedback acts in the opposite direction, by reducing the SST and therefore q , but this takes time due to the heat capacity of the mixed layer, whereas the infrared feedback is effectively instantaneous. Therefore an oscillation is possible, for sufficiently large r . The period of this oscillation, set partly by the mixed layer heat capacity, is in the intraseasonal range and thus coincidentally close to that of the biological oscillations.

[13] For $r > 0.28$, the steady solution is unstable and leads to oscillations without biology. For r not too much larger than this threshold, (e.g., $r = 0.3$ in Figure 2), the addition of biology significantly modulates the variability. For slightly smaller r , ($0.24 \leq r \leq 0.27$), it is stable without biology, but responds strongly to the biologically-induced oscillations, amplifying them. For still smaller r , the biology induces significant oscillations in precipitation (few mm day^{-1} around a mean value) without much amplification of the SST variations, diminishing as r becomes smaller. For very high r and for very low r the biology has little effect on the solutions, whether oscillatory or steady, but the biology has a significant effect over a significant and plausible portion of the parameter space.

[14] Previous studies have demonstrated coupling between biogeochemical processes and upper ocean SST [Siegel *et al.*, 1995; Sathyendranath *et al.*, 1991]. The Western Pacific Warm Pool [Lukas and Lindstrom, 1991; Ohlmann *et al.*, 1998; Lewis *et al.*, 1990] and the Indonesian archipelago are especially sensitive to this effect because heavy rainfall there leads to a shallow mixed layer, so that a significant fraction of the short wave radiation can penetrate below the base of the mixed layer. The size of this fraction is related to the amount of chlorophyll in the water. Hence, variations in chlorophyll

will cause SST variations. It has been speculated in the past that these variations may lead to convection [Siegel *et al.*, 1995]. According to our model, indeed, the relatively small variation in SST which results from changing the vertical profile of light absorption by ocean biota can trigger or modulate intraseasonal variability in the coupled atmosphere-ocean system, amplifying the SST variations initiated by the marine biota through cloud-radiative feedbacks associated with deep convection.

[15] Deep convection in the atmosphere will occur where the SST is highest, usually in the Western Pacific Warm Pool. Our simple model highlights the sensitivity of the climate system to the small variations in SST in the Western Warm Pool that might result from the effect of ocean biota on light penetration. We neglect many physical processes, such as the effect of enhanced stratification on vertical mixing and the effects of variable winds. Moreover, the predator-prey oscillations are not the only way to produce variations in chlorophyll concentration. Forcings that enhance nutrient supply to the surface such as Rossby waves [Strutton *et al.*, 2001; Siegel, 2001] and winds associated with the southeast monsoon probably also play a role. Although long term measurements of bio-optical parameters are available in the central and eastern Pacific [e.g., Chavez *et al.*, 1999], measurements from the Western Pacific that can resolve intraseasonal variability are not available, so the exact cause behind variations in the amount of chlorophyll is still uncertain.

[16] While we do not argue that biological processes are the sole cause of tropical intraseasonal variability, we do suggest that they may play a significant role. While satellite measurements give us good coverage in most parts of the oceans, their view of the surface is blocked in regions of strong convection accompanied by frequent and deep clouds. It is therefore desirable to conduct long term in situ bio-optical measurements in the Western Pacific in order to investigate the coupling between physical and biogeochemical processes in this climatically important region. In addition, this and other recent studies (especially Murtugudde *et al.* [2002]) emphasize the need to consider ecological processes in the context of more accurate climate modeling.

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