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MODELING THE TROPICAL-CYCLONE INDUCED FLUX OF CARBON INTO THE THERMOCLINE

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1. INTRODUCTION

The surface waters of the subtropical gyres are extremely poor in nutrients, with just enough phytoplankton to consume whatever is mixed in. Immediately below the euphotic zone, where it is too dark for photosynthesis, nutrients are plentiful, but through most of the year, there is very little to mix these nutrients up to where they could support life. Tropical cyclones (TCs) are an exception to this rule; the strong wind shears that they bring with them deepen the mixed layer substantially in a very short time. Satellite measurements of sea color have revealed large phytoplankton blooms in their wake, persisting for two or three weeks afterwards until the nutrients mixed into the surface layer are consumed and the phytoplankton that lived on them have died, been remineralized, or sunk (Babin et al., 2004). This process has the potential to form an effective atmospheric sink of carbon. Phytoplankton consume dissolved inorganic carbon (DIC) in the upper ocean and convert it to organic carbon in various forms, some of which sinks out; however, if a DIC deficit is created in the upper ocean, air-sea exchange will allow the upper ocean to absorb enough CO_2 from the atmosphere to make up the difference. There is an invisible effect that may be just as important; when cold water is mixed up from the deep and heated, the solubility of carbon dioxide (and all other gases) decreases, causing an over-saturation and thus an efflux of CO_2. Bates et al. (1998) examined the carbonate chemsitry of three TCs that passed through the Sargasso Sea in 1995 and concluded that the CO_2 efflux induced by them was 55% higher than it would have been in their absence.

A simple but defensible model will be described that resolves the upper column of the ocean, including its biology, from specified atmospheric conditions, along with its parameter dependences and the assumptions that go into forcing it. It will be used to estimate the effect of a single powerful cyclone, and a few limiting factors will be discussed.

2. MODEL PHYSICS

Since the relationship between the physical and biological states of the ocean is set by the behavior of the mixed layer, the model physics of (Price et al. 1986, PWP) are taken, which resolves mixed layer depth from the sources and sinks of energy and momentum, as well as a few stability criteria. The upper 300m of the ocean is resolved at 1m resolution. While numerical stability requires roughly six-hour temporal resolution, some of the inputs are specified at 6-hour intervals, and so a 2-hour time step is used, with inputs interpolated onto a finer grid.

At every time step, the model first applies solar heating to the top several meters of the water column, and, in the uppermost box, applies cooling due to surface fluxes, i.e. outgoing radiation, evaporation, and sensible heat flux. The density of the column is calculated, and a preliminary mixed layer depth is found such that no static instabilities exist. Horizontal momentum is then deposited into the preliminary mixed layer by wind stress, and the depth of the mixed layer is iteratively recalculated such that the bulk Richardson number (Ri_b) is greater than 0.65. (Price et al., 1978) Once the depth of the mixed layer is established, the stability of the submixed layer is calculated, and when shear instability between two layers exists (gradient Richardson number, $Ri_g < 0.25$, following Miles and Howard, 1964) they are allowed to mix together.

Finally, molecular diffusion and Ekman pumping are applied. At the surface, the pumping velocity (w_e) is proportional to the curl of the wind stress. The pumping velocity diminishes with depth; as an approximation, it is assumed that $w_e = 0$ at the bottom of the domain, and that the pumping velocity falls off linearly over the entire column. In order to satisfy continuity, this implicitly means that mass is allowed to flow into or out of the domain from the sides so that the horizontal velocity gradient balances the vertical velocity gradient.

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After all the physical processes have been resolved, the biological processes are allowed to step forward in time. The biological components are treated as passive tracers with regards to the physical model; they are moved with the fluid but do not affect its motion.

3. MODEL FORCING

Because cloud fraction varies day-to-day, and because the daily depth of the mixed layer depends closely on the incoming shortwave radiation, the shortwave fluxes from the NCEP reanalyses were used. (Kalnay et al., 1996) Since the NCEP reanalysis provides a net surface shortwave flux which varies with cloud cover, it is necessary to use the NCEP longwave flux as well. Any attempt to fit the upward longwave flux to the Stefan-Boltzmann law is bound to perform poorly. Prescribing a constant longwave cloud reflectivity would be physically incongruous with allowing shortwave cloud transmittance to vary.

Sensible and latent cooling of the ocean were prescribed with bulk formulae. NCEP fluxes are available for sensible and latent cooling, but using bulk heat transfer parameterizations ensures that the sea surface temperature will always relax toward the atmospheric temperature. While the annual mean NCEP latent and sensible fluxes are fairly close to the mean fluxes arising from turbulent parameterizations, they are not well-correlated on daily timescales. Since a temperature perturbation will be forced on the model (Section 5), this will also ensure that the ocean temperature relaxes back to the air temperature.

Wind forcing presents a thorny problem. While 6-hour winds are available from the NCEP reanalysis, but they frequently disagree substantially with the instantaneous observations from the QuikSCAT satellite. While QuikSCAT observations are prone to their own errors (in particular, accuracy is compromised by rain), they have a stronger physical basis than reanalysis winds. Unfortunately, for a given location they have a poor temporal resolution. This model is forced by QuikSCAT winds, with the stress[1] linearly interpolated between observations and instantaneous observations smoothed by a Gaussian filter with a cutoff of 20 km.

Fresh-water fluxes are neglected, and salinity is taken to be constant throughout the column at all times. This is not an extreme approximation, since the part of the characteristic vertical density gradient caused by salinity is an order of magnitude lower than the part caused by temperature. This reasoning is in not ironclad, since temperature differences shrink in the winter and fresh water fluxes are presumably concentrated into large events (such as TCs), but the ability of the model to recreate observations without including salinity is sufficient that it is not worth the added complexity of adding freshwater fluxes and with them a new external forcing.

4. BIOLOGICAL MODELING

My biological model is very heavily reliant upon Follows et al. (2007). While their model was used to simulate many hypothetical species of phytoplankton in competition for the same resources, a single generic species is modeled here. As with the physics, the goal for the biological model was to use the simplest possible implementation that produced reasonable behavior.

Phytoplankton can be reasonably thought of as converting sunlight (here denoted as PAR, for Photosynthetically Active Radiation[2]), nitrate, phosphate, and carbonate into more phytoplankton, which live, reproduce, and die over a timescale of one to several weeks. Their growth is primarily limited by sunlight and by the availability of nutrients, generally nitrate. Phosphate is not resolved and is assumed to always be present in sufficient quantity to support growth; this is generally the case in the real ocean. Phytoplankton will grow at the rate specified by its more limiting resource.

The dependence of phytoplankton growth on nitrate concentration is expressed in terms of a half-saturation, as proposed by Eppley et al. (1969).

Zooplankton are neglected from the model. Because the flow of material from phytoplankton to zooplankton is a one-way process, and because the goal of this model is to characterize what are essentially short-term perturbations to this flux, this simplification does not meaningfully change the outcome. Instead, phytoplankton will be considered to die and decompose directly into their constituent chemicals at a constant exponential-decay rate. These chem-

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[1] $\underset{\sim}{\tau} = \rho C_D \underset{\sim}{u} \left|\underset{\sim}{u}\right|$

[2] The flux of solar radiation is expressed in fundamentally different units in a biological model than a physical model. PAR refers to the flux of photosynthetically useful photons, not energy. Through this project it is assumed that the surface flux of PAR was equal to the net shortwave flux scaled by a constant ratio of 0.24 MJ = 1 Ein, which has an accuracy of "better than 10%." (Kirk, 1983)

icals will be grouped into Particulate Organic Nitrogen (PON), and Dissolved Organic Nitrogen (DON). Both of these groups are remineralized into nitrate with their own characteristic timescales (λ), but since PON is comprised of solids it falls out with a sinking velocity w_s. This mechanism represents the only way for biological material to leave the system.

Because of the diversity of phytoplankton in the real ocean, the remineralization time and the sinking velocity are very difficult to quantify with a single number. However, *e*-folding depth of PON (z^*) can be measured directly and can be expressed as the ratio of the sinking rate and the remineralization timescale. Worldwide, z^* is found to be about 100-200 (e.g. Schmittner et al. 2005); constants were thus chosen from within the plausible range set forth by Follows et al. (2007) that uphold this relationship.

For notational convenience, phytoplankton concentration will be expressed in terms of the amount of nitrogen tied up in living biota, but the amount of carbon converted into phytoplankton is dictated by the nearly-constant Redfield ratio, which specifies that phytoplankton utilize carbon, nitrogen, and phosphorous at a ratio of 106:16:1. This ratio is not a constant for all species or communities but it holds remarkably well for the phytoplankton community as a whole, so it is sufficient for a model which treats phytoplankton generically. The mechanism that sets the Redfield ratio is not settled (Falkowski, 2000). When phytoplankton grow, they consume carbon species from the seawater, and when their constituent DON and PON are remineralized, the corresponding carbon is returned to the system. In the mixed layer, carbon is also introduced or removed from the system via air-sea exchange, which will drive the system toward saturation, albeit with a fairly long timescale.

The carbon content of the ocean is tracked as total dissolved inorganic carbon (DIC)[3]. The saturation DIC is a function of temperature and the ionic species present, (Follows et al., 2006). However, solving for DIC^* using realistic ocean values shows that it very reasonably be approximated as varying linearly with temperature (e.g. Bakker et al. 1999) experimentally supports the same conclusion. Air-sea exchange of CO_2 brings the mixed-layer DIC into equilibrium on a timescale (λ_{DIC}) of about a year.

In order to initialize the biological variables, nitrate values are initialized to the World Ocean Atlas climatology, Dissolved Inorganic Carbon to the GLODAP climatology, and phytoplankton to a near-infinitesimal value through the whole column, and the dead phytoplankton to be everywhere zero. The full model is then run for 365 days, under repeated diurnal forcing equal to the first day to be modeled, and with Ekman pumping disabled so that the flow of matter into or out of the domain does not interfere with convergence. The final biological state of this spin-up run is then taken as the initial values of my dynamic run.

5. SIMULATING TROPICAL CYCLONE MIXING

In principle, the ocean model is entirely capable of resolving the changes to the mixed layer due to the passage of a storm, given accurate data describing the wind stress and its curl as the storm passes. However, this data is not well-tabulated, and, for most storms, is not even available. The QuickScat instrument, which is used throughout this project, only provides three-day moving averages and non-gridded instantaneous observations (the latter of which are used to force the mixed layer winds), so, short of introducing an enhanced stochastic forcing at the time a TC is known to pass, there is no way for the model to resolve the mixed layer deepening caused by a TC on its own.

Instead, the deepening of the mixed layer is measured more directly. Using the Unisys Best Track database and data from the TRMM satellite, the sea surface temperature (SST) depression is determined by subtracting the average of the SST in the week after the storm passes from the average of the SST in the week before the storm passes. The model is then artificially mixed to a depth such that the surface temperature is depressed by the same amount. In the model, this TC mixing is specified as occurring in the space of one time step, i.e. two hours.

This approximation neglects the cooling due to enhanced surface fluxes when the TC passes; back of the envelope scaling shows that this should be negligible, and Sanford et al. (2007) provide observational data that supports the same conclusion.

Because the same patch will be modeled with and without simulated TC mixing, and because the possibility exists that some part of the TC will be observed by the QuikSCAT satellite (depending on when it passed over relative to the mixing time), the observed winds near the time of TC passage are removed to avoid spurious mixed-layer deepening in

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[3] While the carbon content of the ocean is partitioned between $CO_2(aq)$, HCO_3^-, and CO_3^{2-}, the equilibration time between these is on the order of minutes

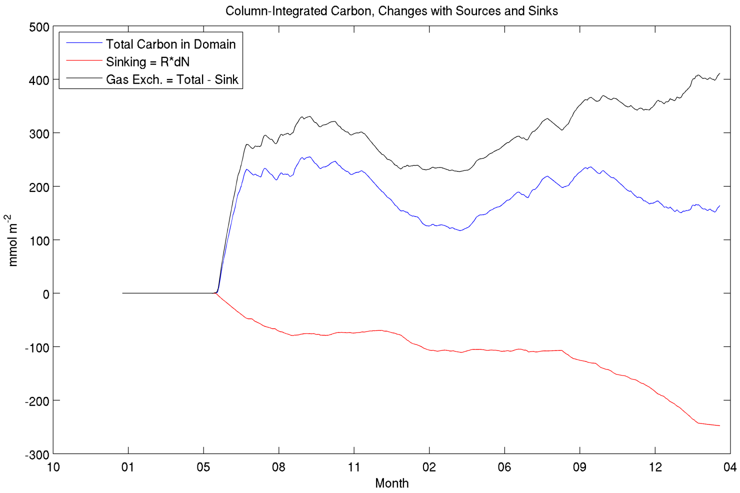

Figure 1: Change in column-integrated carbon over two years, with inferred sources and sinks. The x-axis is time expressed as month of the year, and the y axis is the total carbon in the upper 300 meters, in $\mathrm{mmol\ m^{-2}}$.

the no-TC scenario. This does not have any affect on the run with enforced TC mixing.

6. TYPHOON CHANCHU, 2006

Typhoon Chanchu was a Category 5 storm which passed over the South China Sea on May 15, 2006; it deepened the mixed layer such that the sea surface temperature fell by an estimated 2.5°C. The biological implications of the TC passage appear to be intuitive: a large phytoplankton bloom is induced by the mixing event, and the winter bloom is stronger than in the control run, presumably because the relative warming of the intermediate layer has caused the winter mixed layer deepening to be more pronounced and thus entrain more nitrate-rich water.

It is most illustrative to examine the amount of carbon in the domain, as well as the inferred sources and sinks. A few model characteristics are exploited to facilitate this: first, while carbon can enter or exit the domain both through air/sea exchange and through biological sinking and diffusion at the bottom, nitrogen can only enter or exit the domain through biological sinking and bottom diffusion. However, every mole of nitrogen that sinks out biologically carries R moles of carbon with it (where R is the Redfield ratio).

Figure 1 shows the behavior of carbon integrated over the upper 300 m for the Typhoon Chanchu case minus the same patch of ocean over the same period of time, but with no tropical cyclone mixing. The blue line represents the change in total amount of carbon in the domain (where zero means that there is the same amount in both the TC and the no-TC run), the red line shows the change in nitrogen in the domain multiplied by the Redfield ratio, which shows the amount of carbon removed from the domain due to biological sinking. The black line thus represents the difference between the two, which must come from air/sea exchange. Note that a negative number means that carbon has left the domain due to an inferred process, and a positive number means that carbon has entered the domain. The values plotted represent the total amount of carbon inferred to have entered or left the domain, so the flux can be thought of as the slope of the line.

As we can see from the blue line, there is an increase in total carbon due to the TC immediately after it passes. This is counter-intuitive; naively we might expect that amount of carbon in the domain to decrease because of the mixing of cold, DIC-rich water into the mixed layer, which then must be outgassed as the mixed layer warms. However, because the same mixing entrains a large amount of nitrate into the mixed layer, biological processes quickly create a relative deficit of DIC and the mixed layer quickly becomes under-saturated. In other words, while the mixing event creates a relative surplus of n moles of DIC in the mixed layer, it entrains much more than n/R moles of nitrate, and because biological processes happen on a much faster timescale than air-sea exchange the net effect is that there is less DIC in the mixed layer after the TC than before. Consequently, it is shown (in the black line) that there is a

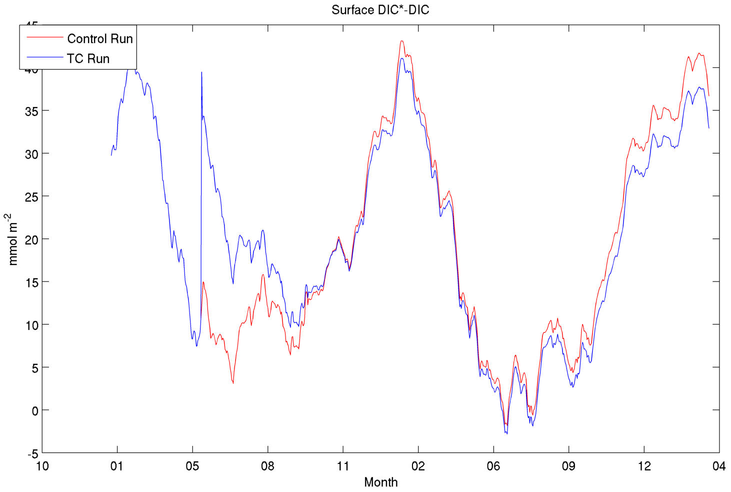

Figure 2: δC modeled for Typhoon Chanchu and the equivalent control. The model produces carbon flux into the ocean for both cases, however the biological processes stimulated by the TC mixing convert mixed layer DIC to biological material, and thus increase the influx of carbon into the domain.

strong flux of carbon into the domain due to gas exchange shortly after the TC, which then essentially levels off.

Sinking of organic particles exports carbon downward at a very roughly constant rate (recall that because the TC run minus the control run is shown, any seasonality has been removed) until, more than a year after the TC event, the vertical PON gradient has re-established itself and there is no more net sinking.

The modeled TC mixing produced a flux of carbon into to upper ocean from the atmosphere, and a flux of carbon out of the domain due to biological sinking; the latter is of greater magnitude than the former and so there is less carbon in the domain at the end of the model period than at the beginning.

An alternative method of characterizing the modeled air-sea carbon flux is to examine the over- or under-saturation of surface DIC, expressed by the quantity $\delta C = DIC^*_{surf} - DIC_{surf}$. It is useful to examine this quantity simply because a larger (positive) δC produces a stronger flux of carbon into the ocean; this gives us a sanity check for the inferred total gas exchange discussed previously; the air-sea exchange values in Figure 1 should be proportional to the time integral of δC. Figure 2 present δC for both the TC and reference case; and show that the gas exchange is qualitatively consistent with what is inferred from the change in column nitrogen; δC rises in the wake of the mixing event, but when it is disabled we see that it falls. The behavior of δC after a TC depends on the change in surface temperature (which changes DIC^*) and the amount of DIC and nitrate entrained (the former directly increases DIC_{surf}, the latter decreases DIC_{surf} via biological processes).

7. SUMMARY AND CONCLUSIONS

The model presented in this paper represents a novel simulation of the physical and biological behavior of the upper ocean in a computationally cheap manner, and has the potential to be used on most storms in the last decade. Given initial conditions for the ocean's temperature, carbon content, and biological state, solar and wind forcing, and the depth of the mixing induced by the passage of a tropical cyclone, this model can describe the effect of the passage of that tropical cyclone on the flux of carbon into or out of the ocean. It has been shown that the characteristic carbon fluxes caused by TC mixing are on the order of several hundred mmol m^{-2}, which is on the same order of magnitude as the background ocean-to-atmosphere carbon flux in the tropics over the course of a year (Bates et al., 1998). However, the net effect of TC mixing is critically balanced between the amount of carbon and nitrogen entrained and the change in saturation DIC due to the cooling of the mixed layer; these effects are of similar size but opposite sign.

This is an impediment toward drawing global, annual conclusions about the role of TCs in the earth's carbon cycle; for example, an attempt was made to verify the findings of Bates et al. (1998) in characterizing the induced carbon flux due to Hurricane Felix (Aug. 1995), which occurred before both the QuikSCAT and TRMM satellites were operational and required a manually-specified SST depression, and it was found that the sign of the net air-sea carbon flux depended on the SST depression. Because the carbon flux depends on the deepening of the mixed layer and the ratio of the entrained nitrogen to cooling, this is unsurprising, however it indicates that extreme caution must be taken to ensure that mixed-layer deepening is properly estimated.

While the model presented could benefit from moving to a three-dimensional framework, which would allow advective processes to be resolved, the most important avenue for future work is the refinement of the model inputs – the carbon-to-nitrogen ratio at depth, the slope of the thermocline, and the mixing depth of a given tropical cyclone. If these are chosen carefully, this model shows great promise of quantitatively illuminating the global effect of TCs on the carbon cycle.

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