Masters Project Report

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Abstract

I proposed a low order model to describe the CO2 flux in the equatorial Pacific and how this varies due to ENSO. The model assumes one contribution due to the temperature of the water and the wind speed at the surface of the ocean, and a second contribution due to water from the deep ocean being drawn up to the surface. Using this approach, I was able to construct a parametrised solution which agrees well with observation on the magnitude of the flux, the difference in flux between the east and the west of the ocean and the inter-annual variation in flux. The model is not however able to capture the North, South variation in flux in this area of the ocean, probably as we have ignored advective currents.

1 Introduction

In this project, I aim to construct a simple model to describe the flux of carbon dioxide between th ocean and the atmosphere in the equatorial Pacific. In particular we would like to understand how the CO2 flux is affected by the El Nino Southern Oscillation (hereafter ENSO).

An El Nino event is when there is anomalously warm water in the equatorial Pacific for an extended period of time, typically several months. The temperature anomaly is usually strongest in the east or the centre of the Pacific basin. The corresponding event with colder than average water is named La Nina. ENSO is the term given to the irregular cycle of El Nino and La Nina event, and has a time period of between 2 and 7 years. (reference)

To understand why ENSO occurs, we put forward an idealised model of the processes which dominate the dynamics in the equatorial Pacific.

Throughout this report, we consider the ocean to be a two layer ocean. This is a simple model in which we describe the ocean as having two distinct layers, an upper, mixed layer, and a lower, deep ocean. For a given point in space and time, each layer is considered to constant properties over its entire depth. Only the mixed layer is able to interact with the atmosphere, and the properties of the deep ocean change only very slowly. On the time period that we are dealing with (the time period of ENSO of a few years) we take the properties of the deep ocean to be constant. (This approach is frequently used in the equatorial Pacific, for example in papers such as (reference).

Under normal conditions, (neither El Nino nor La Nina) the easterly trade winds create a mean westerly flow in the water at the surface of the ocean. In the east of the basin, this causes water from the deep ocean to be drawn to the surface to replace the water moving to the west. We shall call this process of deep ocean water being drawn up to the surface as upwelling. This upwelled water is much cooler, resulting in a cooler sea surface temperature (hereafter SST) in the east of the basin than in the west. We can see the effect of this upwelled water by looking at the depth of the thermocline along the equator. The thermocline is a thin layer in the ocean where the temperature of the water changes very rapidly with depth. In a two layer ocean, this corresponds the surface which marks the boundary between the mixed layer and the deep ocean. Figure() shows that under normal conditions, the thermocline is much shallower in the East of the basin than in the West. We can also see that the during an El Nino event, the profile of the thermocline is much flatter, and is much deeper than before in the East, and so is linked to warmer than average SST here. The reverse is true during a La Nina event, with a steeper profile indicating more upwelling in the East and so lower than average SST.

If we perturb the system away from this mean state, for example by creating a positive SST anomaly in the centre of the Pacific, then we weaken the easterly trade winds (the Bjerkness hypothesis (reference)). We expect that this will reduce the amount of water being upwelled in the East, and indeed it does excite a downwelling wave in the thermocline which will travel towards the East in the form of an equatorial Kelvin wave. At the same time, a westwards moving upwelling Rossby wave is created. We now have an El Nino event, with the thermocline deeper than normal in the East, and shallower than normal in the West. Eventually, the westwards moving Rossby waves will reach the western edge of the Pacific, and will be reflected as upwelling Kelvin waves which cancel the downwelling Kelvin waves and end the event. (Further knowledge ofKelvin and Rossby waves is not required here, except that they propagate with different, well known speeds).

ENSO is the largest observed source of variation in inter annual global SST, and as a result of this, we expect it to also have a large impact on the variability of the CO2 flux between the ocean and the atmosphere. Understanding ENSO, and how it affects the exchange of CO2 is therefore an important step in understanding how CO2 is stored in the ocean. However, ENSO has proved to be a feature that is very difficult to model accurately, with the current climate models struggling to reproduce the correct combination of amplitude, location and power spectrum of ENSO. As a result, predictions of the carbon flux in this region are generally unreliable. Furthermore, due to the complex, numerical nature of these models, it is often difficult to gain any physical insight into the processes occurring.

In light of this, we would like to step back from these complex models, and instead attempt to buld a low order, physically intuitive model of how the carbon flux varies during ENSO. The main goals of this project are therefore to create a model which is as simple as possible while still capturing the main effects, gives an answer which is easy to interpret physically, and is driven by observational data.

2 Experimental Method

In constructing our model, we assume that the two factors which have largest effect on the carbon flux in this region are the surface temperature of the water, and the rate of upwelling. Two notable factors that we have chosen to ignore in simplifying the system are the biological life in the ocean and the horizontal advective currents.

We also assume that it is possible to treat the flux due to the temperature of the water and the flux due to the upwelling as independent effects.

2.1 Flux due to SST

We expect the SST to have an effect on the amount of flux observed as both the solubility of a gas in water and the partial pressure of the gas depend on the temperature of the water. As the water heats up, the solubility decreases, the partial pressure increases, and so we expect a more positive flux at higher temperatures. (reference)?????

We can intuitively expect that the flux of carbon dioxide out of the water will be proportional to the difference in concentration of CO2 between the ocean and the atmosphere. We can therefore write;

$$F_{SST} = k \cdot \Delta[CO_2]. \tag{1}$$

The constant of proportionality here is the gas transfer velocity, and is typically take to be a function of the wind speed at the surface of the ocean only. We will use the form suggested in (reference) of;

$$k(v) = 87.6 \cdot (0.31v^2 - 0.71v + 7.76). \tag{2}$$

It is convenient to present the flux in units of $mol \cdot m^{-2} \cdot yr^{-1}$, requiring k in units of $m \cdot hour^{-1}$ and $\Delta[CO_2]$ in $mol \cdot m^{-3}$. The factor of 87.6 in (2) is due to a unit conversion from $cm \cdot hour^{-1}$. From here, we shall use Henry's constant $H(T)$ which relates for a given temperature, the concentration of gas in a liquid to its partial pressure via;

$$H(T) = \frac{[CO_2]}{pCO_2} \tag{3}$$

$H(T)$ is a constant for a given temperature, so we can obtain a value for it by looking at a simple system with still water and gaseous CO2 at a pressure of one atmosphere above it. The partial pressure is then one atmosphere (approximately the atmospheric pressure P) and the concentration of CO2 in the water is then simply the solubility of the water at this temperature. The solubility of CO2 in water is shown in Figure 2. We approximate the solubility in this temperature range as;

$$S(t) = \frac{1000}{44} \cdot (22.4 - 0.07T) \tag{4}$$

where the numerical prefactor is a unit conversion from $g \cdot kg^{-1}$ into $mol \cdot m^{-3}$ assuming a constant density of water of 1000 kgm^{-3}.

From this we obtain;

$$H(T) = \frac{1000 \cdot (22.4 - 0.07)}{44 \cdot P} \tag{5}$$

Using Henry's constant, we can now write (1) as

$$F_{SST} = H(T)k(v)\{pCO_2^{ocean} - pCO_2^{air}\} \tag{6}$$

where pCO_2 is the partial pressure of CO2. We are now able to use the results of (reference) to parametrise the pCO_2 as;

$$pCO_2^{ocean} = Ae^{0.0423T} \tag{7}$$

where A is a parameter to be chosen later.

Finally, assuming a constant mole fraction of CO2 in the atmosphere, f, gives;

$$pCO_2^{air} = fP \tag{8}$$

where P here is the atmospheric pressure.

This allows us to write the following equation for the flux due to the SST in terms of variables for which there is a large quantity of data.

(perhaps mention the data to get all this in terms of earlier than now!)

2.2 Flux due to Upwelling

We expect the upwelling to have an effect on the CO2 flux because the deep water is far cooler and richer in CO2 than the water in the mixed layer (reference). As this water is brought to the surface, it warms up, and releases some of the CO2 contained within it.

We can quickly estimate the amount of CO2 released per unit area in this process as being equal to the amount of carbon dioxide released per unit volume multiplied by the upwelling velocity, w. We calculate the amount of carbon released per unit volume as the change in the concentration of CO2.

As before, we estimate the concentration as $[CO_2] = H \cdot pCO_2^{air}$ giving us;

$$[CO_2] = \frac{f \cdot 1000 \cdot (22.4 - 0.07T)}{44} \quad (9)$$

which suggests a change in concentration of;

$$\Delta[CO_2] = \frac{70 \cdot f \cdot \Delta T}{44} \quad (10)$$

where ΔT is a parameter representing the change in temperature experienced by the water as it rises.

Finally, this results in a flux due to an upwelling velocity of w of;

$$F_{upwelling} = \frac{70 \cdot f \cdot \Delta T}{44} \quad (11)$$

2.3 Whole Model

Combining these two contributions, we obtain the following equation which will serve as the basis for all of the calculations in this paper;

$$F_{total} = F_{SST} + F_P upwelling need to find out how to do this agai \quad (12)$$

2.4 Estimation of Upwelling Rate

Equation (13) serves as a clear and physically intuitive description of the effects that we will be considering, however, this form of the equation is difficult to use in practice as the upwelling rate w is very small and is almost never measured, and modelled only rarely. This lack of data makes it necessary for us to construct a simple model t approximate the value of w from variables for which data is more available.

We are interested in effects on or very close to the equator, and therefore are considering points which are at almost the sane latitude. Additionally, near the equator we cn assume that currents will predominantly run along the equator rather than across it. These two factors lead us to approximate that if there was no upwelling anywhere else in the basin, we would expect to see vey similar temperature profile at any point in the basin This is not what is observed, and so we make the approximation that the deviations are due to upwelling cold water. A useful variable for us to use here is the heat content of the water per unit area (calculated as an integral of the heat content over the top 300m of the ocean). From our model of the dynamics in the Pacific basin, we expect to see very little upwelling in the west of the basin, and so we assume that the heat content here is what we would expect to find in the absence of upwelling. This will be our reference heat content ($Heat_{reference}$). At all other points, we can then calculate a heat difference, finding that the heat content is lower in the east of the basin. Using our model of a two layer ocean, we can then work out the volume of deep ocean water needed in order to explain the heat difference.

$$\begin{aligned} Heat_{ref} =& \rho c d T_{hot} + \rho c (D-d) T_{cold} \\ Heat_{actual} =& \rho c (d-x) T_{hot} + \rho c (D-d+x) T_{cold} \\ \Delta Heat =& Heat_{ref} - Heat_{actual} \\ & mcx \Delta T \end{aligned} \tag{13}$$

where ρ is the density of water, c, is the specific heat capacity of water, and we have defined $\Delta T = T_{hot} - T_{cold}$.

We now rearrange to find the cold water volume $V_{coldwater}$.

$$V_{coldwater} = \frac{\Delta Heat}{\rho c \Delta T} \tag{14}$$

Finally, we now introduce a parameter, R, the replenishment rate, which represents the frequency with which the cold water volume is replaced.

From this, we calculate the upwelling rate, w, as;

$$w = \frac{R \cdot \Delta Heat}{\rho c \Delta T} \tag{15}$$

2.5 Fitting Model Parameters to Data

Let us summarise our progress so far;

$$F_{total} = Hk \cdot \{Ae^{0.0423T} - fP\} + \frac{70fR\Delta Heat}{44\rho c} \tag{16}$$

We must now estimate the values for the model parameters.

- f: This is the mole fraction of CO2 in the atmosphere. We assume that this is a constant, and has a value of 3.8×10^4 (atmospheric concentration of CO2 is about 380ppm).
- $Heat_{ref}$: This is the heat content that we expect to find with no upwelling. We pick a region in the west of the basin as our reference point, (0N, 156E). We also allow for a seasonal cycle, and o the reference value used will be the average for that month of the year.
- A: This is chosen by picking a temperature and pressure for which the carbon flux due to SST is zero. By looking at Figure 7, we choose a region in the west of the equatorial Pacific as our zero point, and by looking at mean SST and surface pressure in this region, obtained a value of 1.085×10^{-4}.
- R: As there is no precise data for the upwelling velocity, this was chosen to be 400 by tuning the model to best fit the observed flux. The corresponding value for w is peaking at about 5×10^{-4} (assuming a ΔT of $5K$) which is a sensible value.

2.6 Effect of ENSO on Carbon Flux

We now wish to use this model model to calculate how the carbon flux varies with time during ENSO. In order to do this, we need to relate the variables used in equation () to some parameter representing the strength of ENSO at that time. For our purposes, we choose the thermocline depth as our ENSO parameter, (reference) as there is a large amount of data for this, and it is a common output for models describing ENSO. It is a common practice (change wording here) to characterise all variables in terms of thermocline depth, (reference or change) and is convenient for us to take this approach.

For this paper, I use data taken from (reference). The TAO data is gathered from a system of buoys in the Pacific which cover a region from 8N to 8S, and from 137E to 110W. These buys gather data daily, and records go back to about 1960, although gaps in the data are frequent, especially early in the time series. The data set does not include data for the thermocline depth as such, but does include data for the 20C isotherm depth, which we expect to be a surface within the thin thermocline lyer, so we shall use this as a substitute. Due to the frequent gaps in the data, I first convert the data to a monthly mean, thus reducing the impact of individual missing days or weeks. I then calculate a mean value for each of the variables used (SST, heat content, surface air pressure, surface wind speed and 20C isotherm depth). It is important for us to be able to separate the effect of ENSO from the effect of the annual cycle, so we calculate a separate mean for each month of the year. From this, we are able to obtain a time series of the anomaly in each of the variables, calculated by subtracting the expected value for that month from the actual value recorded. We then make a scatter plot of the anomaly in each of these variables against the 20C isotherm depth anomaly. From this, we calculate the Pearson Product-Moment Correlation Coefficient and a line of best fit. The results of this are shown in Table 1 and Figure 4.

We found a strong correlation between the SST and the isotherm depth, and an even better correlation between the heat content and the isotherm depth. The correlations with pressure and wind speed were much weaker, but were still considered good enough for us to gain a rough parametrisation of them from the line of best fit and the monthly mean.

This gives us a method of estimating a carbon flux using the thermocline depth as the only parameter.

2.7 A Simple Model of ENSO

We now wish to demonstrate our ability to generate a prediction of the carbon flux based on a model of ENSO. In this case, we shall use the simple model of ENSO proposed in (reference). This model is excellent for our purposes as it is one of the simplest which still exhibits the low order chaotic behaviour that we observe, has a time period which approximately agrees with observation, and shows seasonal locking. It also characterises ENSO in terms of the thermocline depth alone.

The model is based around the idea of a delayed oscillator. This uses the description of ENSO that we put forward in the introduction, where we have an eastwards propagating Kelvin wave, which reaches the east of the basin a time τ_1 after being created. At this point,the coupling between the ocean and the atmosphere allows for this wave to have an impact on the centre of the basin via the wind. Along with the Kelvin wave, a westwards propagating Rossby wave is also created, and this is reflected off the western boundary of the basin and then travels eastwards as a Kelvin wave. This wave eventually reaches the East of the basin at a time τ_2 at which point it also has an effect on the anomaly at the centre of the basin.

By adding a seasonal cycle to this concept, we can construct a simple equation for the rate of change of thermocline depth;

$$\frac{dh}{dt} = aA[h(t-\tau_1)] - bA[h(t-\tau_2)] + c\cdot\cos(\omega t) \quad (17)$$

where a, b, and c are parameters, and $A[h]$ is a function defining the coupling between the ocean and the atmosphere.

τ_1 and τ_2 can be easily calculated as the time taken for the waves to propagate the required distance;

$$\tau_1 = \frac{L}{2C_{Kelvin}} \quad (18)$$

$$\tau_2 = \frac{L}{2C_{Rossby}} + \frac{L}{C_{Kelvin}} \quad (19)$$

Here, L is the width of the Pacific basin, and C_{Kelvin} and C_{Rossby} are the speeds of the Kelvin and Rossby waves respectively.

We use the form suggested of A in (reference) of;

$$lookuphowtodothisagain!!! \quad (20)$$

The predicted for of ENSO varies greatly depending on the choice of parameters. For a full discussion of how the parameters affect the model, see (reference) and (reference), for our purposes, we shall choose the parameters to be as in (reference) with $\kappa = 2.0$. This choice of parameters gives us the best possible mix of period of oscillation, seasonal locking, and chaotic behaviour. Figure 6 shows a plot of the time series of this model against the observed data.

3 Results

We start by presenting the observed flux in Figure 7. We shall use this to evaluate the success of our model. In Figure 8 we present a plot of the model time averaged carbon flux over the whole of the basin. These plots have been created using Equation 17 with observation data from (reference) to create a carbon flux map for each month since 1979. We not that while there was sufficient data for SST, heat content, isotherm depth and wind speed, the data for pressure does not exist for many of the grid points. In order to preserve enough data points, we have made the assumption that pressure is constant. This is not a terrible assumption as the variance in the surface pressure is small and so ΔpCO_2 is dominated by the pCO_2^{ocean} term. We set this constant to be 101kPa. The time average is calculated for each grid point as a mean of all of the months for which there is data. In all of the plots shown, we have defined a positive value to refer to a flux of carbon dioxide out of the ocean.

We can immediately see that we predict a much larger flux than expected in the North of the basin, but in the South of the basin, our prediction matches the observational data well. Figure 9 illustrates this by plotting the time averaged carbon flux along a slice through the data at a latitude of 4S, where the data suggests that the effect of ENSO is large.

We next wish to examine how the carbon flux varies with time. To do this, we pick a single grid point, in our case (0N, 110W), and plot the time series of the flux at that point. We have chosen this grid point because we showed stronger correlations between the variables in the east of the basin than in the west, and so believe that many of our assumptions will be most reliable in this region. It is also a grid point in which the data is amongst the most complete, and allows for comparison with the model, which is only valid for a point in the east of the basin. Figure 10 shows the plots for the carbon flux as a function of time.

We can also compare how to flux varies in different parts of the basin. Figure 11 compares the flux in the west of the basin to the flux in the east of the basin. Finally, we wish to show that we are able to make a prediction of the carbon flux based on a model of how the thermocline depth evolves with time. In order to do this, we have parametrised all variables in terms of the isotherm depth. We here assess the validity of this method by plotting a comparison between the historical flux based on all of the data and a plot based only on the historic isotherm depth.

We note the the variability in the parametrised model is much lower than in the unparametrised version, but that the mean values and the shape of the plots are similar. Proceeding with this method, we are now able to plot the predicted carbon flux based on the model of the thermocline anomaly. We have scaled the thermocline anomaly by a factor of 50 so that the magnitude of the fluctuations are in line with the observed values. Figure 12 shows the result of this.

4 Analysis

In Figures 10 and 13, we see that our total flux in the East is comprised of a strongly positive flux due to the upwelling, and a slightly negative flux due to the SST. We also note that in general, the anomaly in F_{SST} has the opposite sign to the anomaly in $F_{upwelling}$. This makes sense, as an increase in upwelling results in a decrease in the SST, however, it has an interesting effect on the inter-annual variation of the flux when comparing the parametrised and unparametrised variables (Figure 12). With this parametrised variables, the signs of F_{SST} and $F_{upwelling}$ are always opposite, and so there is always a significant cancellation between them, making the variation in total flux smaller than the variation in either component. This is not true as much in the unparametrised version, where while the signs often opposed, this is not always the case, and the form of the components are not identical. The allows for a much larger variation in the flux.

When we compare the inter-annual variability in the CO2 flux that we obtain from our model with observation, (reference) we find that our variation in the East of the basin roughly agrees with observation, both with a value of about 1 $molm^{-2}yr^{-1}$, but our variability in the West of the basin does not drop off as it does in the observed data. This could be due to the fact that the thermocline is so much deeper in this section of the ocean. This extra depth will mean that the properties of this region will be much slower to change, and could damp the effect of ENSO more than suggested in the model.

From Figure 9, we can see that the model captures the East, West pattern very well. Both the magnitude of the flux and the variation of the flux between the east and west of the ocean agree quite well with observation. This suggests that the proposed model is succeeding in capturing much of the observed behaviour and is an encouraging result. However, our model shows significant North-South gradient which does not agree with the data. A possible explanation of this is the ocean currents. Our weakest assumption appears to be in our estimation of the upwelling velocity, where we have had to attribute all of difference in heat to upwelling. In the North-East of the basin, this is not the case, as there is the additional effect of cooler water being brought down from North of the equator (Figure 14). As a result of this, the water in this region is cooler than elsewhere, and we have over estimated the upwelling rate, giving us the false result. In favour of this explanation, as about 4S, where Figure 14 indicates that the approximation of a purely westwards flow is best, we obtain the results which best match observation.

Another effect that we have ignored in this model is that of biology. We would expect that this would have the effect of decreasing the carbon flux, as the biological life has the net effect of absorbing CO2 (reference) and carrying it down to the deep ocean when it dies. The impact of ignoring this effect is not as obvious in our results, and this is possibly because the amount of biological life is also correlated to the upwelling rate. This is because as well as being rich in CO2, the deep ocean is also rich in other nutrients that help to sustain life. To a first approximation, it may well be reasonable to approximate the contribution from biology as being equal to $const + K_{bio}w$. If this is the case, then the impact of biology would only be to change the values of the parameters used. The *const* term would effect the value of A, and the K_{bio} would effect the value of R. This would suggest that our values for both F_{SST} and $F_{upwelling}$ are slightly too low, and would be difficult to observe in the data. Alternately, the effect of biology could simply be very small in comparison to the other factors.

5 Conclusion

We have shown that it is possible to reproduce many of the features seen in the air-sea CO2 flux in the equatorial Pacific by considering only the flux due to the upwelling of water from the deep ocean and the flux due due to the temperature of the water. We are able to obtain a modeled profile of the flux across the ocean at a latitude of 4S which agrees well with observation both qualitatively and quantitatively. This shows that there is merit in the approach that we have taken. However, it is also clear that there are some areas where the model fails, most severely when the ocean current cannot be assumed to be parallel to the equator. While it might be possible to roughly parametrise the impact of these other horizontal currents, it seems unlikely that they could be easily incorporated into the model in a physically meaningful way. By contrast, further study might be able to extend the model to include the effect of biology, as discussed in the Analysis, however we expect that this would only have a minor effect on the final result.

Finally, it is worth stressing that this model is only valid in the equatorial Pacifi. We have already discussed that our model for estimating the upwelling currents is sensitive to currents that are not parallel to the equator. Even if a model for the upwelling was obtained, there are many other features that this model does not account for, for example, it is not clear how to extend the model to an area of downwelling.