# Masters Project Report

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**Abstract**

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## 1 Introduction

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

We begin by offering a brief discription of some of the terms thats we shall use in the report;

*El Nino:* Anomalously warm water in the equatorial Pacific lasting for several months and normally centred in the east or the centre of the basin.

*La Nina:* Anomalously cold water in the equatorial Pacific lasting for several months and normally centred in the east or the centre of the basin.

*ENSO:* The El Nino Southern Oscillation is the term given to the irregular cycle of El Nino and La Nina events. The Southern Oscillation here refers to the pressure changes between the East and West coast of the equatorial Pacific that accompany these events. ENSO typically has a time period of between 2 and 7 years.

*Two Level Ocean:* Throughout this report, we will aproximate the ocean as a two level ocean. By this, we mean that the ocean is comprised of an upper, mixed layer, and a lower, deep ocean. The upper layer will be warmer, and will interact with the atmosphere. We can also assume that this layer will change on timescales much shorter than a year. By contrast, the deep ocean is colder, and we assume that its properties are constant over the timescales delbt with here.

*Thermocline:* A thin layer in the ocean where the temperature of the water changes rapidly with depth. We can interpret this as a surface which marks the boundary between the mixed layer and the deep ocean.

To understand why ENSO occurs, we first must look at the main proccesses that occur in the Pacific. Under normal conditions, (ie, neither El Nino nor La Nina conditions) the westerly trade winds create a mean westerly flow in both the mixed layer of the ocean and the air above it. In the east of the basin, this causes water from the deep ocean to be drawn to the surface to replace the water which has moved west. We shall call this process of deep ocean water being drawn to the surface upwelling. This upwelled water from the deep ocean is much colder and far more rich in carbon dioxide and nutrients, resulting in a cooler sea surface temperature (hearafter SST) in the east. A further result of this is that the thermocline is much shallower in the east than in the west.

If we now perturb the system away from this, create a rossby wave which propagates westwards, comes back again as a Kelvin wave and reinteracts with the thermocline.

ENSO is the largest observed source of variability found in global SST, and as a result of this, we expect it to also have a large impact on the variability of the carbon flux between the ocean and the atmosphere. Given that the ocean is one of the largest stores of carbon globally, changes in the level of this are important to understand if we are to acurately predict how levels of CO2 will change in the future. One of the reasons why this region is particularly interesting is that the large, complex weather and climate models generally struggle to reproduce the correct combination of amplitude, location and power spectrum of ENSO. As a result of this, predictions based on these models struggle to accurately represent the contribution that the El Nino events represent to the global carbon flux. Furthermore, these models are also in gerneral very difficult to interpret physically.

In light of this, we would like to step back from these complex and detailed models, and instead attempt to build a low order and intuitive model of the carbon flux 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 a answer which is easy to interpret physically, and is driven by observational data.

# 2 Experimental Method

In our model, we assume that the two factors which will have the largest effect on the flux in this region will be the surface temperature of the water and the rate of upwelling in that area. As such, it will be the contribution due to these factors that we use to construct our model. Two notable factors that we have chosen to ignore here are the effect of biology in the ocean and any effect due to advective currents. We have also assumed that we are able to treat these two factors as independant of one another.

## 2.1 Flux due to SST

We expect the SST to have an impact on the amount of flux observed simply beccause the solubility of a gas in water depends on the temperature of the water. As the water heats up, the solubility decreases, and so we expect a more positive flux at higher temperatures.

In our model, we will estimate the CO2 flux due to the temperature of the water as being equal to the difference in the concentration of CO2 in the ocean and the air multiplied by the gas transfer constant. From this, we obtain the following equation;

$$F_{SST} = H(T_{SST})k(v_{wind})\{pCO_2^{ocean} - pCO_2^{air}\} \tag{1}$$

Where $pCO_2$ is the partial pressure of CO2 and $H$ is Henry's constant, which relates for a given temperature, the concentration of a gas in a liquid to its partial pressure via;

$$H(T_{SST}) = \frac{[CO_2]}{pCO_2} \tag{2}$$

$H(T)$ 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 1 atmosphere (101 kPa) and the concentration of CO2 in the water is then simply the solubility of the water at this temperature. The solubility of C02 in water is shown in (figure). We approximate the solubility in this temperature range as;

$$S(T) = 22.4 - 0.07T) \tag{3}$$

Therefore, we finally have that;

$$H(T) = \frac{22.4 - 0.07T}{1.01 \cdot 10^5} \tag{4}$$

The gas transfer velocity, $k$, is typically taken to be a function of the windspeed at the surface of the ocean. We will use the form suggested in (reference) of;

$$k(v_{wind}) = 87.6 \cdot (0.31v_{wind}^2 - 0.71v_{wind} + 7.76) \tag{5}$$

(we have multipiled by 86.7 to convert from units of cm/s to units of m/year)

We use the results of (reference) to parameterise $pCO_2^{ocean}$ as;

$$pCO_2^{ocean} = Ae^{\alpha T_{SST}} \tag{6}$$

Finally, we assume a constant mole fraction of $CO_2$ in the atmosphere, $f$, to give;

$$pCO_2^{air} = fP \tag{7}$$

where P here is atmospheric pressure.

## 2.2 Flux due to Upwelling

We expect the upwelling to have an effect on the CO2 flux because the deep water is know to be far cooler and far richer in CO2 than the water in the mixed layer. 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 in this proccess as being equal to the amount of carbon released per unit volume of water multiplied by the upwelling velocity, $w$. We calculate the amount of carbon realsed as the change in the concentration of CO2.

As before, we estimate the concentration as $pCO_2^{air}) \cdot H$ and so our concentration is estimated as;

$$[CO_2] = fP \cdot \frac{22.4 - 0.07T}{P} = f \cdot (22.4 - 0.07T) \tag{8}$$

Which suggests a change in concentration of;

$$\Delta[CO_2] = f \cdot \Delta T \tag{9}$$

Where $\delta T$ 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 $w$ of;

$$F_{upwelling} = f \cdot \Delta T \cdot w \tag{10}$$

## 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;

$$\begin{aligned} F_{total} &= F_{SST} + F_{upwelling} \\ &= Hk \cdot \{Ae^{aT} - fP\} + f\Delta T w \end{aligned} \tag{11}$$

## 2.4 Estimation of upwelling rate

Equation () 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. This lack of data makes it nessessary for us to construct a simple model to aproximate the value of $w$ from variables for which are measured routinely.

We are interested in effects on or very close to the equator, and therefore are are considering points which are at almost the same latitude. Additionally, near the equator we can assume that currents will predomenantly run along the equator rather than across it. These two factors lead us to assert that if there was no upwelling anywhere in the basin, we would expect to see very similar temperature profile at any point in the basin. This is not what is observed, and we assume 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 a 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 or no 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. 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.

diagram goes here

$$\begin{aligned}
Heat_{reference} &= mcdT_{hot} + mc(D-d)T_{cold} \\
Heat_{actual} &= mc(d-x)T_{hot} + mc(D-d+x)T_{cold} \\
Heat_{difference} &= Heat_{reference} - Heat_{actual} \\
&= mcx\Delta T
\end{aligned} \tag{12}$$

Where we have defined $\Delta T = T_{hot} - T_{cold}$. We now rearange to find the cold water volume $V_{coldwater}$.

$$V_{coldwater} = \frac{Heat_{difference}}{mc\Delta T} \tag{13}$$